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Rocks and Rock Minerals

PHYSICAL GEOLOGY. *By* Chester R. Longwell, Adolph Knopf, *and* Richard F. Flint. Third Edition.

HISTORICAL GEOLOGY. *By* Carl O. Dunbar.

OUTLINES OF PHYSICAL GEOLOGY. *By* Chester R. Longwell, Adolph Knopf, *and* Richard F. Flint. Second Edition.

OUTLINES OF HISTORICAL GEOLOGY. *By* the late Charles Schuchert *and* Carl O. Dunbar. Fourth Edition.

OUTLINES OF GEOLOGY. OUTLINES OF PHYSICAL GEOLOGY and OUTLINES OF HISTORICAL GEOLOGY in one volume.

ROCKS AND ROCK MINERALS. *By* the late Louis V. Pirsson. *Revised by* Adolph Knopf. Third Edition.

GLACIAL GEOLOGY AND THE PLEISTOCENE EPOCH. *By* Richard F. Flint.

Rocks and Rock Minerals

By

Louis V. Pirsson

LATE PROFESSOR OF PHYSICAL GEOLOGY IN THE
SHEFFIELD SCIENTIFIC SCHOOL OF YALE UNIVERSITY

THIRD EDITION

Revised by

Adolph Knopf

STERLING PROFESSOR OF GEOLOGY IN YALE UNIVERSITY

New York: JOHN WILEY & SONS, Inc.
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PREFACE TO THE THIRD EDITION

No new edition of *Rocks and Rock Minerals* has been undertaken since 1925. To bring the contents up to date many changes have been made throughout the entire book and drastic editorial revision has been effected. The treatment and classification of the igneous rocks, it is believed, have been made more rigorously logical. Geologists have not yet shown any great enthusiasm for using a field classification that frankly proclaims itself to be a field classification, but continue to name their rocks in terms of the more refined petrographic systems. Therefore it is unreasonable, for example, that such a term as *diorite* means one thing in a field classification and another thing in a petrographic classification. Such confusion has been avoided by conforming as far as possible to petrographic definitions and by urging the use of purely descriptive terms for field use. I realize that many petrographic and petrologic terms are freely used in areal geologic accounts and in reports on mining districts, and in order that students may be able to read these reports understandingly the more widely used of these terms have been explained.

To take account of the important and difficult class of sedimentary rocks the chapters on those rocks have been rewritten. The treatment of the metamorphic rocks has been brought into accord with modern ideas. Throughout the book more significant chemical analyses of rocks, accompanied by mineral composition and specific gravity determinations, have been freely inserted.

Special pains have been taken to make the definitions define and the explanations explain—and to make them understandable to those not already indoctrinated in the subject. To simplify, clarify, and condense were the guiding principles in making this revision.

Adolph Knopf

Yale University
New Haven, Conn.
October 30, 1946

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PART I. INTRODUCTORY AND GENERAL

CHAPTER I

SCOPE OF PETROLOGY; HISTORY; METHODS OF STUDY

> The aim and end of all geological work is not merely to tell us what rocks are like, but to enable us when we look at a rock to say how and where it was formed. When we can do this, Geology becomes not a mere catalogue of dry descriptions, but a history; and we learn to look upon rocks as the pages of a volume, in which is written an account of what was going on while they were being formed.—A. H. GREEN (1882).

Beneath the mantle of soil and vegetation that covers so much of the surface of the land are rocks, constituting a solid foundation. Here and there, in ledges, in cliffs, and on mountain tops, the rocks jut out and are thus exposed to view. They are visible also in many of the long, deep highway cuts of this era of modern road building. We know too that they must underlie the sea. How far this foundation extends downward we do not know positively, but its lower limit is probably at a depth of 40 or 50 miles. Upon this shell of rock and soil we live and exert our activities; into the rock foundation we penetrate for coal, oil, gas, metals, and the many other mineral supplies essential to our civilization. It is therefore highly important to us, and the information we have acquired in regard to it by examination and study is a valuable branch of human knowledge.

Petrology—the science of rocks. Our knowledge of the constitution and history of the Earth comprises the science of geology. This knowledge has grown so large that geology has split into several well-defined branches, or geologic sciences. Thus geomorphology (or physiography) deals with the surface features of the Earth and how they have been fashioned into our present landscapes. Paleontology is the geologic science based on the study of the remains of past life on the Earth, as presented by the fossils inclosed in the rocks, and it teaches of the manifold forms of life that lived in the geologic past and determines their evolutionary history.

Petrology also has become a separate branch, one of the geologic sciences. It comprises our knowledge of the rocks that make up the Earth's crust, of the constituents of the rocks, of the factors that led to the formation of the rocks, of the laws governing these factors, and of the behavior of the rocks under the action of the geologic agents to which they have been subjected. It endeavors also to classify the rocks into orderly arrangement.

The terms petrology and petrography are not synonyms, although they are often so used. Petrology has already been defined. Petrography concerns itself with the description of rocks and especially with the study of the rocks by means of the microscope as explained later—thus microscopic petrography. Petrography is therefore a branch of petrology. The term petrology is used for the science of rocks as a whole and embraces the geologic, petrographic, chemical, and genetic features of rocks. The synonym lithology is now nearly obsolete.

Definition of a rock. By "rock" is meant the solid material of various kinds that composes the Earth's outer shell. In conformity with idiomatic English, it denotes a hard or firm substance. There are of course border-line cases, where it is difficult to draw a line between coherent, loosely coherent, and incoherent material. Geology recognizes in its terminology the distinction between a rock and the incoherent material from which the rock is derived; for example, *gravel* is the name given to a certain loose material, but if a gravel has been converted during its later geologic history into a coherent mass, that coherent aggregate is given a rock name and called *conglomerate*. Similarly, *till* is the name applied to loose glacial material, but if a till has been changed to a coherent mass, that change is recognized by giving the resultant product a rock name—*tillite*.

In engineering practice the decision as to what is and what is not rock is sometimes highly important. If contracts for excavation are based on rates that vary with the kind of material to be excavated and are higher for rock than for loose material, controversies are likely to arise and the matter may eventually get into the courts. Therefore it is best not to base contracts on terms in which such a twilight zone exists.*

The term rock implies, within limits explained later, a certain constancy in the composition of the mass considered to be a rock of a particular kind. Accordingly, the filling of a mineral vein by an inhomogeneous mixture of quartz, calcite, and ore-minerals is not accepted by petrographers as forming a definite kind of rock.

Composition of rocks. A rock is sometimes defined as an aggregate of one or more minerals. This definition is not broad enough, however, for although most rocks are made up wholly of minerals, some consist

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* F. L. Ransome, What is rock? *Econ. Geol.*, vol. 28, pp. 502–505, 1933.

entirely of glass, and some are a mixture of minerals and glass. A mineral is a substance that not only has a definite chemical composition but also has as a rule a crystalline structure. Glass is a substance that was once molten but was chilled so rapidly that it solidified without any of its potential minerals being able to form. A rock composed wholly of minerals may be either a monomineral or a polymineral aggregate; that is, a rock may consist of one kind of mineral alone, as, for example, some of the purest marbles, which consist of calcite only. More generally a rock consists of a mixture of two or more minerals, as for example granite, which is made of grains of quartz, feldspar, and mica. These subjects are treated in detail in later chapters.

History of petrology. The science of geology can be said to have commenced when rocks, as objects of investigation, began to be studied. As the individual minerals that make up the rocks or are contained in their cavities were investigated by chemical means and by the goniometer, the science of mineralogy and its related subject, crystallography, began. At first the difference between rocks and minerals was not clearly perceived. Fine-grained rocks composed of mineral grains so minute that they cannot be distinguished by the unaided eye or by the magnifying lens were thought to be homogeneous substances and similar in their nature to minerals. This baffling state of affairs continued even down to the middle of the last century.

As the knowledge of the composition and properties of minerals grew, the coarser-grained rocks were found to be composed of grains of one or more kinds of minerals, and early in the last century many common rocks had already been named in accordance with their component minerals. As investigation proceeded and the science of geology grew, many new rocks were discovered and named, and thus the list of named rocks increased. It should be said that the early geologists, armed only with a simple lens, were exceedingly keen observers and made many surprisingly correct observations on the mineral composition of fine-grained rocks. Schemes of classification were proposed, some of them containing admirable features; but the extremely fine-grained rocks defied the means of investigation then at command, and their composition, properties, and relations to other rocks remained perforce largely unknown.

About the middle of the last century, Sorby, a brilliant English geologist, showed that by a suitable technique very thin slices of rocks (approximately 0.001 inch thick) can be prepared. By examining such thin sections under an appropriate microscope he demonstrated that the component mineral grains can be determined and that their properties, their relations to one another, the order in which they were formed, the processes to which they had subsequently been subjected, and many other interesting and important features of rocks can be discovered. He showed that this method could be successfully applied to the finest-grained rocks. This method of investigation was immediately taken up with enthusiasm, especially in Germany by Zirkel and others. A new and brilliant era in the study of rocks began, and the science of petrography grew rapidly. A flood of knowledge regarding rocks, and especially the minerals composing them, began to rise and has kept on rising to the present day. The study of the properties of transparent minerals under the action of polarized light received a great impulse, and the facts discovered have in turn been of immeasurable service in investigating rocks by this method.

Classification of rocks. According to their mode of origin rocks fall into three main groups. These are the *igneous* rocks, made by the solidification of molten material; the *sedimentary* rocks, formed by the deposition of sediments in water, to which are added the small groups of eolian, or wind-formed, and ice-formed deposits; and the *metamorphic* rocks, produced by the transforming action of certain powerful geologic processes acting on either igneous or sedimentary rocks. In metamorphic rocks the original characters have been wholly or partly replaced by new ones, and the resultant rocks are therefore most conveniently considered separately. The three-fold grouping is used in the present work, and each group is treated by itself with its subdivisions.

Summarizing what has just been said, we have:

I. Igneous Rocks: solidified molten masses.
II. Sedimentary Rocks: deposited detrital material and precipitates.
III. Metamorphic Rocks: formed by the transformation of rocks of groups I and II.

Field and petrographic classifications. The use of the microscope in the study of thin sections has shown that rocks apparently absolutely identical, either in the field or in hand specimens, may be composed of entirely unlike minerals. Chemical analysis also has proved that apparently similar rocks may differ fundamentally in chemical composition. Such rocks are therefore quite distinct, deserving separate names and places in any classification in which all the essential characters of rocks are considered, and yet outwardly there may be no hint of this difference to the eye. Two useful terms, *megascopic* (Greek, *megas*, great) and *microscopic*, are employed: megascopic refers to those characters of the rocks that can be perceived by the eye unaided or aided by a simple pocket-lens, and microscopic to those that require the use of the microscope on thin sections. Manifestly, a classification that is based as much on microscopic traits as on megascopic ones cannot be used in determining rocks in the field. It is more precise, but in the nature of things it cannot be of general use. This subject is treated more fully in the section devoted to igneous rocks. Suffice it to say here that this book attempts to supply a field classification that is based on the megascopic characters of rocks, determinable by the eye or pocket-lens, aided by a few simple means for determining the minerals. Many important facts that do not depend on microscopic study have been discovered regarding rocks, in recent years, and it is intended also to give some account of these discoveries.

Microscopic petrography. Although this volume is not concerned with the microscopic method of research, a brief sketch of how such research is conducted and the sort of results obtained by it is of interest. To prepare a thin section of a rock for study, a chip of the rock as thin as can be obtained is taken. It should be, for the normal thin section, about an inch in diameter and of firm, unaltered material. It is first ground flat on a metal plate with coarse abrasive powder and water and is then ground smooth on another plate with very fine powder. The smooth sliver is cemented to a piece of glass by means of heated Canada balsam, and the other side is ground down with the coarse emery until the section is as thin as cardboard, or as far as this operation can safely be carried with the coarse powder. It is then in a similar way ground down with finer abrasive and finally finished on a glass plate with the finest abrasive flour until, if it is a dark, fine-grained rock, it becomes so thin that ordinary print can be read through it. It is next enveloped in balsam, and a thin cover glass is placed over it; it is then ready for use. The professionals who make a business of preparing such "thin sections" save much time in the coarser work by the use of circular saws charged with diamond dust or by using carborundum powder on disks or endless revolving wires. They become highly expert in the final grinding and prepare sections that are quite uniformly about 0.001 inch thick, experience having shown that this thickness is best for examining the thin section under the microscope. The appearance of one of the finished thin sections is shown in Fig. 1. In such a thin section even the minute mineral grains that make up the finest-grained and blackest of basaltic lavas become transparent and can be determined under the microscope.

The microscope used in petrographic work differs from that ordinarily employed in that it is equipped with an apparatus arranged for studying thin sections in polarized light. Underneath the stage of the microscope is a nicol prism of calcite through which the light passes in coming upward from the reflecting mirror below; as it passes through the nicol it becomes polarized; that is, the vibrations of the light, instead of occurring in all directions, as in ordinary light, are reduced to a definite direction in one plane. As a result of this arrangement, the thin section is viewed in polarized light.

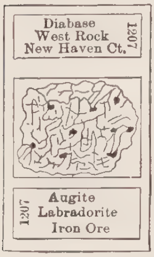


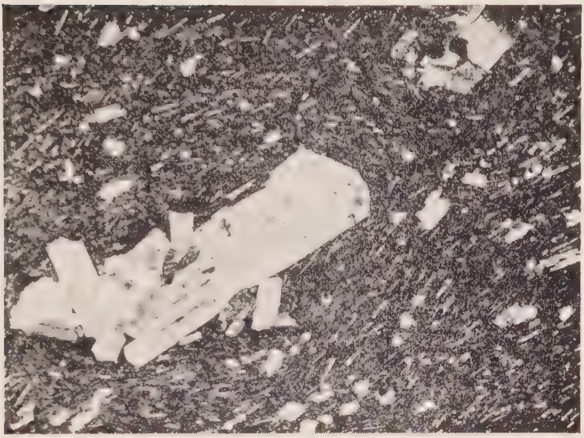
FIG. 1. Thin section of rock.

Another nicol prism, called the analyzer, is fitted in the tube above the object lens so that the effects produced by the mineral particles on the polarized light as it passes through them can be tested and studied. The nicols can also be removed, so that the effects on ordinary light in passing through the minerals can be seen. Other arrangements are provided for strongly converging the light as it passes through the minerals and for testing, in a variety of ways, the results produced by the convergent light.

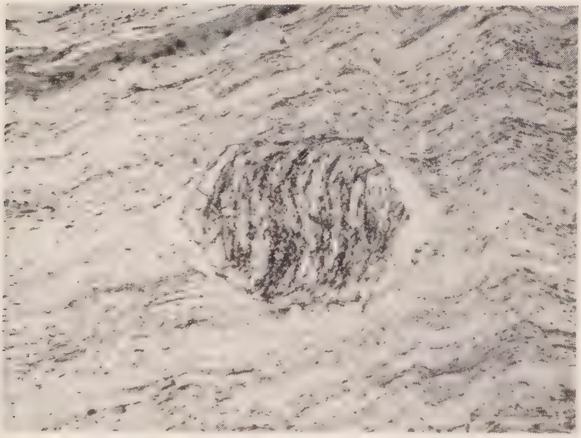
Subjected to such investigation, the transparent minerals of the rocks show many distinctive properties by means of which they can be easily and definitely determined. Crystals or fragments of crystals of almost incredible minuteness can be studied under high powers of the microscope, their properties can be examined, and the particular variety of mineral to which they belong can be made out. In order to use this means of studying rocks, a good knowledge of general mineralogy, crystallography, optics, and in particular the optical properties of the rock-making minerals, is essential. Because of the cost of the apparatus and the technical knowledge required in its use, this method of studying rocks can never become a popular or general one, but many of the results that have been attained by it are easily understandable and can be presented.

Results of microscopic research. By the method just described it is possible to determine the kind of mineral or mineral grains that make up the most compact and fine-grained rocks; to tell whether the rock is of sedimentary or igneous origin; and, if it is of igneous origin, to determine the general order in which the several minerals have crystallized from the molten fluid. One can see whether the crystal particles are clear and homogeneous or whether they contain inclusions of various kinds—facts that may throw light on their origin and history; whether they are fresh and unchanged or have been altered by weathering and changed wholly or partly into other substances; whether they have been subjected to the enormous pressures of moun-

PLATE 1.



A. Andesite Lava (Magnification 23x).
(H. A. Brouwer.)



B. Garnetiferous Schist (Magnification 30x).
(U. S. Geological Survey.)

THIN SECTIONS OF ROCKS AS SEEN UNDER THE MICROSCOPE.

tain building in the crust and are strained and crushed. It is possible to tell at once if a rock has more or less glass associated with its component mineral grains and, if it has, thus to learn with certainty its igneous origin.

Furthermore, if the grains are not too microscopically fine it is possible not only to identify the minerals present, but also to measure their areas or diameters in a given section, to ascertain from this the relative amounts of the different minerals, and then, if the chemical compositions of the component minerals are known, to compute the chemical composition of the rock as a whole. Chemical composition is often of great value in scientific classification and in other ways.

These are some of the more significant features of rocks that can be discovered by microscopic study, and they suffice to illustrate the great power of the method in aiding geologic research.*

Megascopic study of rocks. Although rocks require the microscope for their complete investigation, many important features can be observed without its use. In the coarser-grained rocks, those in which the grains are 1 millimeter† or more in diameter, the component minerals can generally be identified by the aid of the lens or by simple means. Even if much finer-grained than 1 millimeter, some minerals can be recognized by certain of their characters, as explained in the chapter on the rock-making minerals. Even if a rock is so fine-grained that its component grains cannot be distinguished from one another, the color, the hardness, the style of fracture under the hammer, the specific gravity, and the behavior of fragments or of the powdered rock under the action of acids are important characters that serve to distinguish it.

Implements and apparatus. The first requisite is a suitable hammer for obtaining material. It should be a square-faced geologic hammer.

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* The following works may be mentioned in which rock-making minerals and rocks are treated and classified, in a more or less detailed way, according to the results of microscopical research. A. N. Winchell, *Elements of Optical Mineralogy,* Fifth Edition, John Wiley & Sons, New York, 1937. A. F. Rogers and P. F. Kerr, *Optical Mineralogy,* McGraw-Hill Book Co., New York, 1942. H. Rosenbusch and A. Osann, *Elemente der Gesteinslehre,* Fourth Edition, 1924. A. Harker, *Petrology for Students,* Seventh Edition, Cambridge University Press, 1935. J. P. Iddings, *Igneous Rocks,* John Wiley & Sons, New York, 1913. F. F. Grout, *Petrography and Petrology,* McGraw-Hill Book Co., New York, 1932. F. H. Hatch and A. K. Wells, *The Petrology of the Igneous Rocks,* George Allen & Unwin, Ltd., London, 1939. F. H. Hatch, R. H. Rastall, and Maurice Black, *The Petrology of Sedimentary Rocks,* George Allen & Unwin, Ltd., London, 1938. Alfred Harker, *Metamorphism,* Second Edition, Methuen & Co., Ltd., London, 1939.

† One millimeter = $\frac{1}{25}$ inch.

It is convenient to have one end wedge-shaped. The steel should be as hard as possible without being too brittle; otherwise the edges wear off very rapidly. If the hammer is made to order, the hole should be as large as possible, consistent with strength, and tapered somewhat toward the top; the handle, when thrust through the eye of the hammer, will be smaller than the outer part of the hole, and needs to be enlarged with iron wedges. This device, commonly used in securing the handles of picks, is a great convenience, as it prevents the head coming off when the handle shrinks. The hammer should be of good weight, about 2½ pounds for the head, to enable good-sized pieces of rock to be readily broken up and the fresh material within to be procured. Of course, anything in the way of a hammer or sledge may be used on occasion, but this implement will give the best service for general use.

If, in addition to obtaining material for examination, it is desired to trim and shape it into specimens for a permanent collection, a small trimming hammer will be found convenient. It should be double-headed, square-faced, and of very hard steel, and the head should weigh about 6 ounces. Hand specimens for collections should be trimmed to the standard size of about 4 × 3 × 1 inches. They are made by selecting a suitable large spall obtained by the large hammer and knocking small chips from it along the edges, first on one side and then on the other, until it is trimmed to the required shape and size. A well-made specimen shows no hammer marks ("stone bruises"), or at most only on the edges and none on the faces.

A pocket-lens also is essential; one of the apochromatic triplets made by several makers of optical instruments is best, because it gives a flat field, but much cheaper ones will serve the purpose. A lens of focal distance of 1 inch, *i.e.*, having a magnification of 10 diameters, is the most useful for general purposes. Those of higher magnification have too small a field.

Besides the above, which are for use in the field, a small amount of the apparatus that is used in the laboratory for determining minerals is often of service. This would include a blowpipe and platinum-tipped forceps for testing fusibility; pieces of quartz, calcite, and ordinary window glass for testing hardness; a simple apparatus for determining specific gravity; a magnet; a few test tubes, dilute acids, and a Bunsen gas burner or alcohol lamp for testing solubility; and a glass funnel, filter paper, and a few reagents, such as solutions of ammonia, silver nitrate, and barium chloride, for making tests by chemical reactions. A small agate or steel mortar is needed for grinding fragments of rocks and minerals to powder for making chemical tests. This list might be increased almost indefinitely into the full equipment of a mineralogical laboratory. Most chemical laboratories contain all the apparatus and reagents necessary for determining the minerals and rocks mentioned in this book, and if such a laboratory is not available the material mentioned covers nearly all necessary demands and can be used almost anywhere.

CHAPTER II

CHEMICAL COMPOSITION OF THE EARTH'S CRUST AND ITS COMPONENT MINERALS

Composition of the Earth's interior. The origin and history of the rocks composing the outermost shell of the Earth are bound up with the origin and history of the globe itself. Beyond the history that is revealed in the rocks, however, our ideas on the earlier stages of the history of the Earth must, with our present knowledge, be largely speculative. The view most generally held is that the Earth was once a molten mass, the outer portion of which as the result of cooling solidified to a solid crust. No remnants of the crust thus formed—the Earth's "original crust"—have yet been found. The outermost rocky shell of our planet, for convenience called "the crust," is of very different origin from the "original crust." The rocks that make up the crust as it is conventionally spoken of are legion; they have been formed by the activity of geologic forces working and reworking the material of the outermost shell during the long history of the Earth, which is regarded as exceeding 2000 million years.

Facts that are known. On the other hand, much is known of the Earth beyond what we can see at the surface. We know, for instance, that the temperature increases downward in the crust. We know also that there are bodies of molten material, which, though small as compared with the size of the Earth, are nevertheless large, and we see in active volcanoes impressive manifestations of these molten masses at the Earth's surface. We know that such molten bodies not only exist in the Earth's interior now but also were present in past geologic ages, as shown by the fact that molten matter has been forced upward into the crust or has been erupted on the Earth's surface many times during its long history. We know that wherever the deep-seated rocks, which underlie all the stratified and metamorphic rocks, have become exposed to our view by erosion they present the general characters of igneous rocks, and we therefore infer that they were once molten. As the sedimentary and metamorphic rocks have been derived from rocks that anteceded them, this fact leads us logically to infer that they were derived from rocks originally like the deep-seated rocks and that their substance at some time in the geologic past had been in a molten state.

Rock material, then, having been wholly or at least very largely in a molten state, manifestly we should know something of the nature and properties of the molten fluids from which it was formed. We can obtain some knowledge of these molten liquids at active volcanoes, where some of their properties are shown, but the properties most important in the forming of rocks can best be determined by studying the igneous rocks formed by the solidification of the molten masses. This subject is therefore treated in the chapter on them. However, certain aspects can well be considered here, and one of these is the general chemical composition of the Earth's crust.

Chemical composition of the Earth's crust. Nearly 40,000 chemical analyses have now been made of rock specimens from the Earth's crust. Most of them are analyses of rocks from Europe and the United States, but enough have been made of specimens from the rest of the world, in conjunction with microscopical studies of other specimens, to permit some conclusions of wide application. One of the most important of these is this: the average chemical composition of the Earth's crust underlying the continents is everywhere similar.

The statement thus broadly made needs explanation. It does not mean that any one portion of the crust of the continents is composed of the same chemical elements in exactly the same proportions as any other portion. It means that the crust is made up of the same elements and that, although these differ greatly in proportions from place to place or from one kind of rock mass to another, if we take large areas underlain by many kinds of rocks the average composition of such areas is very nearly alike. Thus the composition of the average rock computed from all the analyses made of specimens from the United States is essentially the same as the average computed from the analyses of the rocks of Europe. The composition of the average rock of New England is essentially that of the Rocky Mountains region. On the other hand, much of the Province of Quebec is composed of rock of one kind that extends with monotonous sameness over a vast area; the composition of this rock differs greatly from that of the average rock, and if we were considering this particular part of the continent we should have to increase the area greatly to obtain a correct average. Limestone, which is essentially carbonate of calcium, covers parts of the continental areas, but it is a relatively thin coating.

The average compositions of the exposed igneous rocks of many countries and of the "average igneous rock" for the whole world have been computed. The average composition of the sedimentary rocks has also been computed. By combining this average with that of the average igneous rock in the proportions assumed to hold down to a depth of ten miles, the average composition of the "ten-mile crust" has been computed. The most recent figures are given in Table 1.

TABLE 1

COMPOSITION OF AVERAGE IGNEOUS ROCKS

	A *Average Igneous Rock, United States* [1]	*B* *Average Igneous Rock, World* [2]	*C* *Average Composition of Outer Ten-Mile Crust* [3]
Oxygen	47.34	46.60	49.13
Silicon	28.40	27.72	26.00
Aluminum	8.31	8.13	7.45
Iron	4.53	5.00	4.20
Calcium	3.60	3.63	3.25
Sodium	2.50	2.85	2.40
Potassium	2.19	2.59	2.35
Magnesium	1.78	2.09	2.35
Hydrogen	0.14	0.13	1.00
Titanium	0.44	0.44	0.61
Carbon	0.03	0.03	0.35
Chlorine	0.05	0.05	0.20
Phosphorus	0.12	0.08	0.12
Sulfur	0.05	0.05	0.10
Manganese	0.12	0.10	0.10
All others	0.40	0.51	0.39
	100.00	100.00	100.00

[1] Based on A. Knopf, *Jour. Geol.*, vol. 24, pp. 620–622, 1916.

[2] V. M. Goldschmidt, *Norske Vid. Akad. Oslo, Skr.* 1. *Math. Nat. Kl.*, 1937, No. 4, pp. 99–101, 1938.

[3] R. C. Wells, *U. S. Geol. Survey Bull.* 878, pp. 3–5, 1937.

The average composition of the igneous rocks exposed at the Earth's surface in the United States is given in column *A*. It is a weighted average, obtained by taking into account the sizes of the exposed areas of the various igneous rocks. The composition of the average igneous rock for the world, shown in column *B*, which is not weighted according to areas, does not differ significantly from that calculated for the United States. Column *C* purports to give the average composition of the outer ten-mile crust and includes in proper proportions the igneous rocks, the sedimentary rocks, the water of the oceans, and the gases of the atmosphere; even so, the figures do not differ greatly from those in columns *A* and *B*. Vast as are volumes of the sedimentary, water, and gaseous envelopes of our planet, they are inconsiderable in comparison with the volume of igneous rocks. The figures obtained are first approximations to the truth, and accordingly are of value in showing the relative proportion of the chemical elements in the outer part of the Earth. They have aroused great interest in widely diverse fields of science.

Recent investigations tend to show that the deeper portions of the crust, such as are revealed in the areas of Precambrian rocks, approximate the composition of a granodiorite or granite carrying 68 per cent of SiO_2. Finland is the only large area so far for which the composition of the exposed crust has been accurately determined.

The elements of major geologic importance. In Table 1 the first eight elements are seen to be present in abundance. Oxygen forms one-half of the outer part of the Earth, and the quantity in the atmosphere and in the ocean is small compared with that locked up in the rocks. Silicon comes next, forming about one-fourth of the crust, and then come aluminum (8 per cent) and iron (5 per cent). They are followed by calcium, sodium, potassium, and magnesium. After these come four elements, which, though far less abundant, demand memtion; they are hydrogen, titanium, carbon, and chlorine. Of these, titanium plays no important part in geologic processes. Hydrogen and carbon, however, are of much importance; they are highly active in geologic processes. All the other elements, however important in special cases, either as vitally essential for living organisms or as industrially useful, are from the standpoint of general geology of minor interest.

Combinations of chemical elements. The elements, with few exceptions, do not occur native, or in the uncombined state, in nature. They are in chemical union in some form, comprising the compounds known as minerals. In terms of chemistry we say that they are either combined with oxygen as oxides or are in combinations as salts.

From what has just been said, the following oxides are clearly of chief petrologic importance. We deal primarily with the oxides of the elements, because by custom the results of chemical analyses of minerals and rocks are reported in the form of oxides.

Acid Oxides

SiO_2, silica; in combination and free as solids.
CO_2, carbon dioxide; in combination and free as a gas.

Basic Oxides

Al_2O_3, alumina; in combination and free as a solid; sometimes acid-forming.
Fe_2O_3, ferric oxide; in combination and free as a solid.
FeO, ferrous oxide; only in combination in solids.
MgO, magnesia; in combination and free as a solid.*
CaO, lime; only in combination in solids.
Na_2O, soda; only in combination in solids.
K_2O, potash; only in combination in solids.

* The solid MgO occurs as the mineral periclase but is rare.

The list gives, from top to bottom, the order of decreasing acidity and increasing basicity. To the list we should add also water (H_2O), which occurs free in the gaseous, solid, and liquid states and occurs also in combination.

Since we are dealing with rocks, we need to consider only those oxides and other compounds that form solids. These solids, in descending order of abundance, are (1) silicates, (2) carbonates, and (3) the oxides of silicon, aluminum, and iron.

Ice, the solid form of water, can be regarded as a rock, but as such needs no further consideration in this work.

PART II. ROCK MINERALS

CHAPTER III

IMPORTANT PROPERTIES OF MINERALS

As all rocks, with few exceptions, are composed of minerals, it is highly important to an understanding of them to have a good knowledge of the chief rock-making minerals. This knowledge is indispensable. Consequently Part II of this book is devoted to the rock-making minerals, and the study of the rocks themselves is deferred. Part II contains, first, a general account of the properties that aid in determining minerals megascopically; second, in a succeeding chapter, descriptions of the minerals individually; and third, methods for determining the minerals.

Minerals defined. A mineral is any substance that has been formed by geologic processes and has a chemical composition expressible by a formula. To this definition, for petrographic purposes, should be added that a mineral is a solid and generally has a definite crystalline structure. The term "mineral" is commonly used in two ways: in one, which can be called the mineralogic sense, we mean a compound that has a certain composition and crystalline state, as in speaking of calcite we mean the crystallized compound $CaCO_3$, the rhombohedral carbonate of calcium; in the other usage we speak of "the minerals of a rock" and mean thereby the individual constituent grains, the distinct entities that make up the rock.

Crystals defined. Most chemical compounds, when their atoms or ions are free to arrange themselves in space and conditions are proper for them to assume solid form, as for example when they separate out from solution and grow as *crystals*. That is, the constituent ions arrange themselves in the growing compound in a definite geometric pattern, which is characteristic of that compound and is governed by mathematical laws, and which gives the resulting crystal a definite internal structure and an outward form bounded by planes that make definite, characteristic angles with one another. Thus minerals crystallize into cubes, octahedrons, prisms, etc. The conditions prevailing while a rock is forming may be such that a growing mineral can assume its characteristic outward form, and the resulting solid is a crystal bounded by distinct planes. More commonly, however, adjacent growing crystals interfere with one another and consequently are without distinct crystallographic form. In sedimentary rocks, the constituent mineral grains are chiefly fragments of former crystals whose plane surfaces have been worn off by attrition. All irregularly shaped grains are called *anhedrons* (from the Greek, meaning "without planes").

List of properties. The chief properties by which the rock-making minerals can be recognized are *crystal form, color, cleavage,* and *associations.* These properties can be perceived by the unaided eye. In addition there are *hardness, specific gravity,* and the behavior *before the blowpipe* and with *chemical reagents,* properties that require the use of apparatus to be determined.

CRYSTAL FORM. The mineral grains that compose rocks do not as a rule, as already mentioned, have good outward crystal form. The reason for this is that in the igneous and metamorphic rocks the growing minerals interfere with one another's development, and thus, although they may roughly approximate the general shape that they would have had if they had grown unhampered, the outer surface does not consist of smooth definite planes; in the sedimentary rocks the grains are either broken fragments or have been rounded by rolling and grinding. However, when crystallization begins in a molten liquid minerals of one or more kinds may begin to grow at widely scattered nuclei and complete their growth before the later-crystallizing minerals have begun to grow. Here the crystals are not interfered with, and consequently they are likely to have good outward crystal forms, bounded, therefore, by distinct planes. Such well-formed crystals are excellently shown in those igneous rocks described elsewhere in this book as *porphyries.* Likewise, certain minerals in metamorphic rocks commonly occur in prominent well-shaped crystals, which indicate that the conditions were such as to permit them to develop freely their own characteristic crystal faces while they were growing. Striking examples are the excellent crystals of garnet, often seen in the rock known as *mica schist.*

In general, distinctive crystal form is wanting in minerals inclosed in rocks, and it cannot be used as a means for determining them. In special cases, however, it may be well developed in rock-making minerals and it can then be very useful. The shapes in which each mineral is most likely to occur are described under the heading of that mineral in the descriptive part of this book.

COLOR. The color of minerals, if used with caution, is a useful property for helping to distinguish them. Color may depend on chemical composition, when it can be said to be *inherent*, or it may be due to some foreign substance distributed through the minerals and acting as a pigment and can be termed *exotic*. Because the color of many minerals is exotic, caution must be used in employing it as a means of identification.

In regard to *inherent* colors, neither silica nor carbonic acid in combination as silicates and carbonates have any power to produce color, and so far as they are concerned, such compounds would be colorless, or, as is presently explained, white. So, silica alone, as quartz, is colorless. The same is also true of the metallic oxides, alumina, lime, and magnesia, and of the alkalies, soda and potash. Thus, carbonate of calcium (calcite), carbonate of calcium and magnesium (dolomite), oxide of aluminum (corundum), silicates of calcium and magnesium, and silicates of aluminum and the alkalies, or feldspars, are all inherently colorless. The metallic ions that chiefly influence the color of rock minerals are those of iron, chromium, and manganese, and the only one of these of wide petrographic importance is iron, especially iron in the ferric state. Minerals that contain iron as a prominent component are dark green, dark brown, or black, and these colors can ordinarily be regarded as indicative of this metal.

The *exotic* colors of many minerals are due to one of two causes. A minute amount of some compound having an intense coloring power may be present in chemical combination in a mineral. Thus a minute amount of manganese oxide in quartz is supposed to produce the amethyst color, traces of chromic oxide color some silicates green, and probably copper does the same. Again, the color may be due to a vast number of minute particles dispersed through the crystal as inclusions. These minute specks may have a distinct color of their own and thus act as a pigment, as when, for example, quartz is colored dark red by a reddish dust of ferric oxide particles in it; or the inclusions may be so arranged in regular systems as to act refractively upon light, breaking it up and producing a play of prismatic colors, or opalescence, in the substance. Usually in the latter case, one color predominates and determines the color of the mineral. A good example is the feldspar called labradorite, a constituent of the rocks *anorthosite* and *gabbro*; this feldspar may show a fine play of colors, a rich dark blue usually predominating.

The white color of many rock-making minerals may be due to minute inclusions, as when feldspars inclose innumerable minute scales of kaolinite or white mica; but more commonly it is due to the reflection of light from the surfaces of countless microscopic cracks that everywhere traverse the mineral substance. Such minerals are really colorless and transparent. The effect of the cracks is the same as if a piece of colorless glass were ground to powder, which would of course be white. Hence, white minerals are not regarded as having any color. If free from cracks, they are colorless and transparent. Good examples of this are such common minerals as quartz, calcite, and feldspars. As explained in the discussion of cleavage, the cracks in some feldspars are so regularly arranged as to produce a play of colors, giving the mineral an opalescence or pearly luster and a distinctly predominating color tone like that mentioned above as produced by inclusions.

STREAK. In determining minerals we can be aided not only by the color they show in the solid form but also by the color they show when reduced to powder. The powder can be obtained by grinding a small fragment in a mortar, but it is more easily produced by scratching a sharp point of the mineral across a plate of unglazed porcelain; the color of the resulting streak is, of course, that of the powdered mineral. Although any piece of unglazed porcelain will serve fairly well, small plates—"streak plates"—specially made for this purpose are sold by dealers in chemical apparatus.

The color shown by minerals in the powdered state is generally much lighter than that which they display in the mass and in fact for some minerals is a different color. The streak is most useful in helping to discriminate the dark-colored minerals, especially the metallic oxides and sulfides of the heavy metals used as ores. Hence application of the streak test to the light-colored silicates and carbonates that comprise the rock-making minerals is much less helpful. In these minerals it is sometimes useful in distinguishing exotic colors from inherent colors; for the color of the streak is generally that of the mineral substance itself, and the pigment or other impurities that produce an exotic color must be present in very large amount to exert a perceptible influence on the color of the streak. Thus calcite is generally colorless or white, but some varieties are yellow, brown, or red. The streak of all these varieties is white or barely tinted, except in unusual instances. Feldspars are normally white or colorless, but in some rocks, such as *anorthosite*, they may be black and can easily be mistaken for an iron-bearing mineral; they give a white streak, however, which discloses their real nature.

In the field, the bruised surface of a rock where struck by the hammer generally shows patches of the powdered minerals, thereby giving in a rough manner the color of the streak; or a bit of a mineral can be ground between two hammer surfaces, and the powder can be rubbed on white paper.

CLEAVAGE. If minerals have crystalline structure, the arrangement of the atoms composing them is generally such that the force of cohesion is less in some particular plane or planes than in others. Along such planes, if suitable means are employed, such as placing the edge of a knife on the mineral and striking with a hammer, the mineral tends to split or cleave. Some minerals, such as mica, which is used for stove windows and lamp chimneys, are capable of being split almost indefinitely into thin leaves; others, such as the feldspars, have a good cleavage; but some, as quartz, have no apparent cleavage.

If the cleavage is very good, the new surfaces are smooth and shining like the original surfaces of a crystal, and the cleavage is called *perfect*. The property of cleavage, being so distinctive, is of great use in helping to determine minerals. As already indicated, the directions of cleavage do not occur at random in a mineral; on the contrary, they are definitely related to the special crystal form that characterizes a particular mineral. If the mineral has two systems of planes along which it can be cleaved, like feldspar, for example, the angle between the two planes is, for a feldspar of a certain definite chemical composition, always the same. Some minerals, such as mica, have only one plane parallel to which the cleavage is good. Others have two systems of cleavage; in some minerals the two cleavages are exactly alike in ease of splitting, but in others the two are unlike, one being better than the other. Again, there may be three differently oriented planes parallel to which cleavage can be produced, all alike in ease of splitting as in calcite, or unlike, as in barite ($BaSO_4$); or there may even be four or more. Whether the cleavages are alike or unlike when there is more than one, depends not only on their direction in the crystal, but also on the geometric form or system in which the mineral has crystallized. Description of these relations involves too much of the principles of crystallography to discuss in this place, but the following will be helpful to understand certain terms frequently used.

A. Good cleavage parallel to one plane only: the mineral grains in the rock tend to occur as tablets, folia, or scales, the surfaces of which are parallel to the cleavage. This tendency is well shown in such minerals as the micas and chlorite.

B. Good cleavage parallel to two systems of planes and both alike in ease of splitting: the minerals are likely to occur in elongate forms parallel to the cleavage, which is called *prismatic*. Elongate forms determined by prismatic cleavage are shown by the minerals hornblende and pyroxene. If the two cleavages are not exactly alike, the mineral still is generally elongate in the direction of the edge produced by the meeting of the two cleavage planes. It may sometimes be columnar and sometimes tabular, parallel to the better cleavage. The feldspars, which form the free-developed crystals in porphyry, generally show such tabular or columnar forms.

C. Good cleavage in three systems of planes and all alike in case of splitting: if the three planes are at right angles to each other the mineral breaks into cubes and the cleavage is *cubic* or apparently so; if they are at some other angle, rhombs will be produced and the cleavage is called *rhombohedral*. Cubic cleavage is well shown by galena (PbS), the common ore-mineral of lead, and by rock salt; it is not shown by any common rock-making mineral. Rhombohedral cleavage is characteristic of the common rock-making carbonates—calcite ($CaCO_3$) and dolomite [$CaMg(CO_3)_2$]. Three unlike directions have the same practical effect as two unlike, and four directions are not of importance in megascopic petrography, as no common rock-making mineral shows them.

If a rock made up of mineral grains sufficiently large to be readily studied by the aid of the pocket-lens—the size of peas, for example—is carefully examined, the cleavage surfaces of the minerals will be found almost without exception to be full of minute cracks and fissures. These cracks are parallel at least to one cleavage and generally to all the cleavage directions the mineral has. Besides the cleavage cracks, there are also irregular lines of fracture, which do not correspond to any definite direction. Commonly, the mineral grains of rocks not only contain these large cleavage cracks and irregular fractures that can be perceived with the eye or with the aid of the lens but, in addition, are everywhere rifted by similar ones so minute that they can only be detected in thin sections of rocks under high powers of the microscope. The reflection of light from these minute microscopic cracks renders many minerals opaque and white that would otherwise be colorless and transparent. These cracks and fissures were produced by the various forces to which the rocks were subjected; some are due to the contraction during cooling from a heated stage, as in metamorphic and igneous rocks, and more generally to the intense pressures and strains to which many of the rocks of the Earth's crust have been subjected. Minute as the rifts in the mineral grains are, they are of great importance in geologic processes, for by means of them, and drawn by capillary action, water containing dissolved carbon dioxide penetrates not only the rocks but even the interiors of the individual grains, and alters them into other minerals and changes the rocks into soil.

Fig. 2. Conchoidal fracture in obsidian (volcanic glass).

Fracture. The appearance of the surface obtained by breaking a mineral in a direction other than that of cleavage, or by breaking a mineral that has no cleavage, is called the *fracture* of the mineral. If the mineral is fibrous in structure, the fracture may be *fibrous;* or it may be rough and uneven, or *hackly;* if the mineral is fine-grained and homogeneous it will be *conchoidal,* that is, it will present a smooth, shell-like appearance such as is shown on surfaces of broken glass, which recall the inside of a clam shell. Rocks that are extremely fine grained and homogeneous, such as some flints and glassy or fine-grained compact lavas, also have a more or less pronounced conchoidal fracture. Quartz is the most common mineral that gives a good conchoidal fracture.

Associations of minerals. That certain minerals are likely to occur together in rocks of a certain kind, and that the presence of certain minerals excludes the presence of certain others are facts of practical use. Thus quartz and nepheline are incompatible and consequently they do not occur together in the same rock. This fact of their incompatibility is of value in distinguishing between certain rocks. The important relationships of this kind are mentioned in their appropriate places.

Hardness. Hardness is of great value as an aid in identifying minerals, and it is likewise useful in the field in making rough tests of rocks. The hardness of a mineral is determined by comparing it with the hardness of the minerals in the following scale.

Scale of Hardness

1. Talc
2. Gypsum
3. Calcite
4. Fluorite
5. Apatite
6. Feldspar
7. Quartz
8. Topaz
9. Corundum
10. Diamond

This scale means that each mineral, if used in the form of a sharply pointed fragment, will scratch smooth surfaces of all the minerals preceding it in the list but of none succeeding it. If, for example, a fragment of an unknown mineral is found to scratch calcite, its hardness is greater than 3; if it will not scratch fluorite but on the contrary is scratched by it, its hardness is less than 4, and must be between 3 and 4, or approximately 3½.

The point of a pocket-knife blade as ordinarily tempered with a hardness of little over 5 and pieces of common window glass with hardness of about 5½ are useful for testing the hardness of common minerals and of the rocks made up of them. A common brass pin point is a little over 3 and can scratch calcite; the fingernail is a little over 2 and can scratch gypsum.

Specific gravity. The specific gravity of a substance is its density compared with water, or the number of times heavier a given volume of the substance is than an equal volume of water. It is obtained by weighing a piece of the mineral or rock in air and then in water; the difference between the two is equal to the weight of an equal volume of water (the volume displaced), and we have

$$\frac{\text{Weight in air}}{\text{Weight in air} - \text{Weight in water}} = \text{Specific gravity}$$

The operation can be carried out with one of the special forms of apparatus devised for determining specific gravity and described in the manuals of determinative mineralogy, or it may be done with a chemical, an assay, or a jeweler's balance. The specimen is first weighed in the pan and then suspended from it by a hair and weighed in water. A piece about ½ inch in diameter is convenient for minerals, though generally a fragment smaller than this is selected in order to obtain pure homogeneous material; without such purity, it is perhaps needless to say, the determination of specific gravity is of little value. Adherent air bubbles and air in cracks are best eliminated by boiling the fragment in water and then allowing it to cool before weighing.

If a mineral has an invariable chemical composition, as, for example, quartz (SiO_2) or calcite ($CaCO_3$), its specific gravity has a fixed value, such as 2.65 for quartz, and departure from that value must be due to the presence of impurities. Most minerals, however, are solid solutions and therefore vary considerably in chemical composition. Thus minerals at one end of a series contain magnesium and at the other end ferrous iron, and between these extremes all degrees of mixture of the two. In accordance with such variations the specific gravity of the mineral has a range of values. The pyroxenes, amphiboles, garnets, and olivines exemplify such ranges of specific gravity, and solid solution accounts for most of the ranges in specific gravity of the individual minerals in Table 1 (page 24).

Blowpipe reactions. Most rock-making minerals, which are chiefly silicates and carbonates, do not show before the blowpipe characteris-

TABLE 1

Specific Gravities of Rock Minerals
Arranged in descending order

5.26	Hematite	2.9–3.1	Biotite
5.17	Magnetite	2.87	Dolomite (pure)
5.0	Pyrite	2.80	Talc
4.4–4.9	Ilmenite	2.76–3.0	Muscovite
4.02	Corundum	2.76	Anorthite
4.0–4.4	Goethite	2.71	Calcite (pure)
3.75	Staurolite	2.69	Labradorite
3.6–4.0	Limonite	2.65–2.75	Chlorite
3.5–4.3	Garnet	2.65	Quartz
3.4–3.65	Topaz	2.62	Albite
3.56–3.66	Kyanite	2.6	Kaolinite
3.35–3.45	Vesuvianite	2.5–2.65	Serpentine
3.2–3.7	Pyroxene	2.619	Nepheline (pure)
3.27–3.37	Olivine	2.54	Orthoclase (pure)
3.25–3.45	Epidote	2.45–2.50	Leucite
3.24	Sillimanite	2.32	Gypsum
3.16–3.2	Andalusite	2.28	Analcite
3.15	Apatite	2.18–2.25	Natrolite
3.1–3.47	Hornblende	2.15–2.2	Heulandite
3.1–3.2	Chondrodite	2.14–2.40	Sodalite
3.0–3.47	Amphibole	2.1–2.2	Stilbite
2.96	Anhydrite	1.98	Montmorillonite
2.9–3.0	Tourmaline	1.9–2.3	Opal

tic reactions by which they can be readily determined, as so many of the oxides and sulfides of the heavy metals do. Still, the relative degree of *fusibility* shown by thin splinters, *the coloration of the flame,* and the characters of the *beads* formed by fusing the minerals with a flux are properties that help in determining these minerals. To the extent that these properties are determinative aids they are mentioned in the descriptions of the minerals. If instruction in the use of the blowpipe is desired, it can be found in one of the manuals devoted to that purpose.

Chemical reactions with reagents. Certain qualitative chemical tests that can be made with a few reagents and simple apparatus are of great help in determining minerals and in classifying rocks. In Chapter V, where the methods for identifying minerals are given, these tests are fully described.

CHAPTER IV

THE ROCK-MAKING MINERALS

Section 1. SILICATES AND OXIDES

Silicates and oxides comprise the most important rock-making minerals. They form the chief minerals that crystallize out from the molten fluids risen from the Earth's interior and are consequently the main components of igneous rocks. They make up also in large part the metamorphic rocks and are abundant in sedimentary rocks.

A. SILICATES

Silicates were formerly regarded as salts of various silicic acids in which the hydrogen atoms have been replaced by metals or radicals composed of metals in combination with oxygen, hydroxyl, and fluorine. The three silicic acids thus considered to form the rock-making minerals are *polysilicic acid* ($H_4Si_3O_8$), *metasilicic acid* (H_2SiO_3), and *orthosilicic acid* (H_4SiO_4). Analysis of the framework of minerals by means of X-rays has shown, however, that there are no silicic acids and that the fundamental structural unit in all silicates is a tetrahedral group (SiO_4) in which a central silicon ion is linked with four oxygen ions.

The silicates abundant as rock-making minerals fall into a few groups: the feldspar, feldspathoid, mica, pyroxene, amphibole, olivine, garnet, epidote, and chlorite groups. These groups of minerals are solid-solution series, in which the composition ranges between certain limits determined by definite laws. Any particular mineral in any of the groups has a composition that is determined by the conditions under which it was formed. Practically alone among rock-making minerals, quartz has an invariable composition; all other rock-making minerals are members of solid-solution series.

FELDSPARS

The term "feldspar" designates a group of minerals that have a general similarity in physical and chemical properties. The feldspars are indeed so much alike in appearance and general properties that they cannot be told apart megascopically except in specially favorable circumstances. They are therefore best treated as a group; however, the distinctive characters will be pointed out by which, if possible, they can be told apart megascopically.

The rock-making feldspars comprise three pure feldspars and two sets of solid solutions.

PURE FELDSPARS

Orthoclase, $KAlSi_3O_8$, silicate of potassium and aluminum.
Albite, $NaAlSi_3O_8$, silicate of sodium and aluminum.
Anorthite, $CaAl_2Si_2O_8$, silicate of calcium and aluminum.

SOLID SOLUTIONS

Alkali feldspars, $(K, Na)AlSi_3O_8$, solid solutions of orthoclase and albite.
Plagioclase feldspars, $(NaAlSi_3O_8)_x + (CaAl_2Si_2O_8)_y$, solid solutions of albite and anorthite.

The pure feldspars, i.e., those having nearly the composition indicated by the ideal formulas of orthoclase, albite, and anorthite, are found mainly as crystals lining druses in veins or as crystals rooted in the walls of the cavities known as miaroles that occur in many igneous rocks (page 131), and to a subordinate extent they occur as rock-making constituents.

Most rock-making feldspars are more complex in composition. They fall into two groups: (1) *alkali feldspars* (a convenient term to express that they are feldspars containing the alkali elements potassium and sodium), and (2) *plagioclase feldspars.* The alkali feldspars comprise chiefly a series of solid solutions of orthoclase and albite in varying proportions, the orthoclase generally predominating. Many of these solid solutions have become "unmixed," so that much of what seems to be homogeneous orthoclase is in reality a microscopic aggregate of orthoclase and albite called microperthite. The plagioclase feldspars comprise a continuous solid-solution series that ranges from pure albite to pure anorthite. Albite, it appears, has a dual role: it is both an alkali feldspar and the first member of the plagioclase series.

Because the plagioclase feldspars are frequently mentioned in geologic literature, it is desirable to indicate how they are named. Any plagioclase can be expressed by the formula Ab_xAn_y, where Ab stands for albite, and An stands for anorthite, and $x + y = 100$, measured in molecular percentages. Plagioclase feldspars ranging from Ab_{100} to $Ab_{91}An_9$ are collectively termed albite; those ranging from $Ab_{90}An_{10}$ to $Ab_{71}An_{29}$ are termed oligoclase; and so on. More broadly, plagioclase in which the albite component predominates is called *sodic plagioclase,* and that in which the anorthite component predominates *calcic plagioclase.*

Form. Orthoclase is monoclinic, and, if in well-formed crystals, it commonly has the forms shown in the accompanying figures. Some crystals are stout and thick as in Fig. 3; some are thin and tabular parallel to the side pinacoid *b*, as in Fig. 4; and others are long and columnar, as in Fig. 5. In orthoclase the basal pinacoid *c* is at right angles to the side pinacoid *b*. In albite and anorthite, both of which are triclinic, the faces *c* and *b* are not quite at right angles but are slightly oblique. This is true also for all of their mixtures, that is, for the plagioclase group as a whole. Some mixtures of orthoclase and albite, as well as certain varieties of the pure potassium compound $KAlSi_3O_8$, called *microcline,* are also slightly oblique, but in all the feldspars mentioned the departure from a right angle amounts to only a few degrees and, even under favorable conditions, can scarcely be perceived by the eye and must be measured by a goniometer to be appreciated. It cannot, therefore, under ordinary circumstances be used as a means to distinguish between the alkali and the plagioclase feldspars. The forms of the crystals in which the plagioclase feldspars appear in rocks when they are well crystallized are like those shown for orthoclase in Figs. 3 to 5.

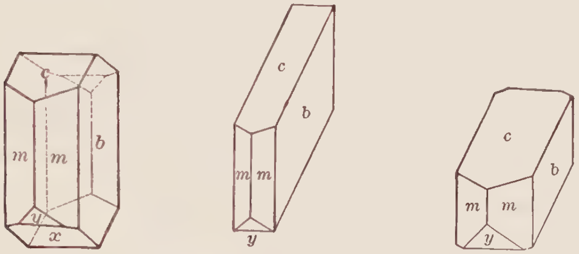

Fig. 3. Fig. 4. Fig. 5.

Orthoclase.

Only as phenocrysts in porphyritic igneous rocks and as projecting crystals in miarolitic cavities in granular igneous rocks have these minerals had the opportunity to assume the free crystal forms described; ordinarily they were interfered with during crystallization by the growth of adjacent minerals or of other crystals of feldspar, and consequently they are in irregularly shaped grains. Nevertheless, feldspars tend to assume these free crystal forms, and in some rocks, such, for instance, as the syenites, which are composed mainly of feldspar, they may have more or less perfectly the shape of flat tablets or of rude laths, according as they approximate to Fig. 4 or Fig. 5.

Twinning. Crystals frequently appear to be compound, as if built of two halves, one of which has been revolved 180° around an axis, perpendicular as a rule to the plane of junction along which the two parts have grown together. A mineral so built is called a twinned crystal. Feldspars are generally twinned, one of the most abundant kinds of twinning being that illustrated in Fig. 6, known as *Carlsbad twinning,* from the town of that name in Bohemia where excellent examples have been found. It is as if a crystal like that shown in Fig. 3 were cut through parallel to the side pinacoid *b*, one of the parts were revolved 180° around a vertical axis parallel to the edge *mb*, and the two parts were pushed together so that they would mutually penetrate. In Fig. 7 the same arrangement is shown as seen by an observer looking at the side pinacoid *b* face of the crystal; *acya* is the outline of the original crystal; if this is cut out in a piece of paper and then turned over 180°, or upside down, and laid on *acya* so that the edges *aa* are brought together, it will give the result seen in the figure. In the twinned crystal illustrated in Fig. 6 the basal pinacoid *c* slopes toward the observer, the dome *y* slopes towards the rear; in the other half this arrangement is reversed. As explained under the cleavage of feldspars, this fact is of importance in helping to recognize these twins when the outward crystal form is imperfect or wanting. Carlsbad twins are found in all the feldspars; they are best developed in the phenocrysts of porphyritic igneous rocks, especially in the large orthoclase phenocrysts of some granitic porphyries.

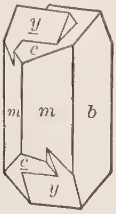

Fig. 6.
Carlsbad twin.

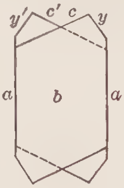

Fig. 7.
Side view of Carlsbad twin.

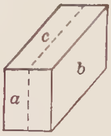

Fig. 8.
Carlsbad twin, showing position of twinning seam on *a* and *c*

In a Carlsbad-twinned crystal the plane of junction of the two parts is parallel to the side pinacoid *b*; the axis on which one part is revolved is the vertical line parallel to the edge *ab* of Fig. 8 and not one perpendicular to *b* or parallel to the edge *ac,* differing thus from the usual mode of twinning. The basal pinacoid *c* in orthoclase makes a right angle with the side pinacoid *b*; the outline of the face *a* is, therefore, a rectangle, and if the crystal were divided along the dotted line by a plane parallel to the side pinacoid *b*, and one of the halves were revolved 180° on an axis parallel to the edge *ac*, that is, perpendicular to *b*, it would appear precisely as before and no twinning would occur. The crystallographic reason for this is that the side pinacoid *b* is a symmetry plane, since the crystal is monoclinic, and a symmetry plane cannot be a twinning plane.

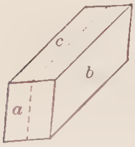

Fig. 9.

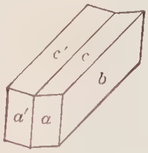

Fig. 10.

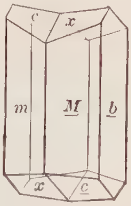

Fig. 11.

Twinning in plagioclase.

In the plagioclase group—albite, anorthite, and intermediate members—the basal pinacoid makes an oblique angle with the side pinacoid; the front pinacoid *a* is, therefore, a rhomboid and not a rectangle, as shown in Fig. 9. If this crystal is divided along the dotted line and one of the halves is revolved 180°, it will present the appearance seen in Fig. 10; the face *c* and the lower *c′*, now brought on top, slope toward each other, forming a re-entrant angle, while below they produce a salient angle. A twinned crystal is therefore produced. Twinning of this kind is known as *albite twinning,* because it is so prevalent in albite. A complete crystal of this kind is seen in Fig. 11. The crystallographic reason for this twinning being able to occur is that these feldspars are triclinic; they have, therefore, no symmetry plane, and any one of the faces may serve as a twinning plane.

Multiple twinning. In actual practice, we rarely find in nature a single albite twin of the kind described above. In the rock-making plagioclases the crystals are divided again and again into thin slices, and these are alternately twinned upon one another, producing the *multiple twinning* shown in Fig. 12. Indeed, the albite twinning descends to such a remarkable degree of fineness that twins may be less than 0.0001 inch thick and can scarcely be perceived even in thin sections in polarized light under the highest powers of the microscope.

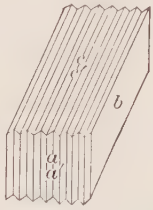

Fig. 12. Multiple albite twinning.

Frequently, however, especially in the more calcic feldspars, such as labradorite, the twinning is coarse enough to be readily seen by the unaided eye: the basal cleavage surface of such a twinned feldspar appears as if a series of fine parallel lines, or striae, had been ruled on it, as illustrated in Fig. 13. Even if the striae are exceedingly fine lines and occur on a small cleavage surface of a feldspar grain embedded in a rock, they can be distinctly seen by properly adjusting the light reflected from the cleavage surface.

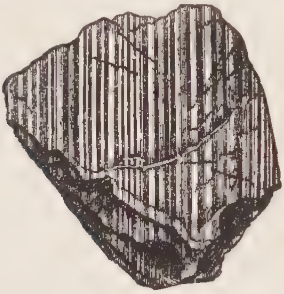

FIG. 13. Broad albite twinning lamellae in calcic plagioclase.

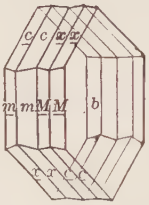

FIG. 14. Albite twinning combined with Carlsbad twinning.

Many plagioclase feldspars are twinned according to both the Carlsbad and the albite laws. Such feldspars can be seen to be divided into the Carlsbad halves by the reflection of light from the basal cleavage, and each of these halves is seen to be striated by fine lines resulting from multiple albite twinning. The combination of these two modes of twinning, whereby each Carlsbad half is divided into albite twins, is shown in Fig. 14. The practical use of twinning is explained in the paragraph on methods for determining the feldspars. Other kinds of twinning besides those mentioned occur in the feldspars, but in the megascopic study of rocks they are not important.

Cleavage. All the feldspars are alike in having two good cleavages, the better one parallel to the base and the other parallel to the side pinacoid (see Fig. 5). Since in orthoclase these two faces intersect at a right angle, so also do the cleavages, and from this fact the mineral takes its name (Greek, *orthos*, straight, right, *klasis*, a break or fracture.) In the plagioclase feldspars from albite to anorthite, these faces are slightly oblique, and so are the cleavage planes; hence the name plagioclase (Greek, *plagios*, oblique, *klasis*, a break or fracture) was given to the group.

In rocks, if the feldspar grains are of good size, the cleavages can readily be seen by reflected light; they are commonly interrupted, giving rise to steplike appearances. Even if the grains are small, the cleavage can usually be detected with a lens in good light. If the feldspars are more or less altered, as described under alteration, they have lost more or less completely their capacity for showing good cleavage faces on broken surfaces, and this fact must be taken into account in making determinations. In cleavage pieces, as in crystals that show distinct faces, the amount of obliquity of the plagioclases is far too small to be used in distinguishing them by the eye or lens from orthoclase, which is right angled.

On a fractured rock surface, if the crystal grains are sufficiently large, the cleavage surfaces show that many of the grains are twinned according to the Carlsbad law. A grain thus twinned appears to be divided into two roughly equal parts by a distinct line, called the twinning seam. On one side of this line, if it points away from the observer, the cleavage surfaces slope away in one direction; on the other side of the line they slope toward the observer at an equal angle, like the two *c* faces in Fig. 6, to which in fact they are parallel. This difference in the cleavage directions in the two Carlsbad halves can usually be seen by shifting the position of the crystal in a good light until the reflections from the cleavage surfaces are obtained. At the same time, if examined with a good lens, the cleavage surfaces generally show the closely spaced, parallel striae of albite twinning, which accordingly indicate that the feldspar grain is a plagioclase. The albite twinning striae are parallel to the Carlsbad twinning seam.

Fracture. In directions in which they do not cleave, the brittle feldspars break with uneven and somewhat conchoidal fracture.

Color, luster, and streak. Feldspars have no inherent color; hence, as explained under the color of minerals, they should normally be either limpid and colorless, or white. Transparent, colorless, glassy feldspars in rocks are confined to fresh and recent lavas, in which they commonly occur as conspicuous crystals ("phenocrysts"); they practically do not occur in massive granular rocks like granites and gneisses. The luster in fresh lavas may be strongly vitreous. More commonly, feldspars are semi-translucent or opaque and white, grayish white, or yellowish, and of a somewhat porcelain-like appearance.

Orthoclase and microcline (the potassium feldspars) generally have a tinge of red, ranging from a pale flesh tint to a strong brick red or brownish red; a distinct flesh color* is most common. It is this color in the feldspars that gives many granites used for building stones their color. Most probably the red color of the potassium feldspars

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* A pale pink mixed with a little yellow.

is caused by finely disseminated particles of ferric oxide, which acts as a pigment, and it must be considered as an exotic and not an inherent color.

Plagioclase feldspars are commonly gray, rarely red. They differ from the potassium feldspars apparently because of a difference in the chemical behavior of iron toward sodium and potassium: sodium readily combines with iron in silicate minerals, but potassium does not. Thus in potassium feldspars the iron tends to be present as free oxide and to color the mineral. Therefore, many rocks containing potassium feldspars tend to be reddish, whereas those containing sodic feldspars tend to be gray. This distinction can be used to some extent to indicate what kinds of feldspar are present, but it is not an absolute rule, because many potassium feldspars are white or gray, and, conversely, the plagioclase feldspars in some rocks are red. In general, if in a rock containing feldspars of two kinds, one is red and the other is not, the red feldspar is almost certainly a potassium feldspar (orthoclase or microcline).

Some potassium feldspars, especially the variety of microcline called amazonstone, where occurring in distinct crystals in the miarolitic druses of granites, have a green color—pale to bright grass green. This color is probably an exotic effect produced by a minute amount of an unknown constituent.

Some rock-forming feldspars are gray, bluish gray, violet brown, or even black. The dark colors are more common in the calcic plagioclase feldspars, especially labradorite. They are caused by minute black particles disseminated through the feldspars, which act as a pigment and which may be magnetite dust but are much more commonly ilmenite. Fine examples of dark feldspars are seen in the labradorite rocks from Canada, the Adirondack region of New York State, and Labrador, which are called anorthosites.

Other feldspars have a pearly bluish opalescence from innumerable minute cracks regularly arranged, which reflect the light so that interference colors are produced.

The *luster* of feldspar is vitreous, rarely glassy (glassiness is limited chiefly to the variety called sanidine). Feldspars that are more or less altered commonly have a waxlike appearance and a waxy luster; if completely altered they are dull, chalky, and lusterless.

The *streak* is white and is not characteristic.

Hardness. Feldspar has a hardness of 6; it is scratched by quartz, scratches glass, but is not scratched by the knife.

Specific gravity. The specific gravity of ideally pure orthoclase corresponding to the formula $KAlSi_3O_8$ is 2.539; albite is 2.62; and anorthite is 2.76. The specific gravity of the actual minerals, which are mixtures and solid solutions, ranges between these limits. Thus the alkali feldspars, which consist of a mixture of orthoclase and albite, average about 2.57. The plagioclase feldspars, constituting a continuous series of solid solutions between albite and anorthite, have specific gravities that range from 2.62 to 2.76, depending on the relative amounts of albite and anorthite. If the specific gravity of a fragment of feldspar can be determined with accuracy to the second place of decimals, it affords a fairly good method of ascertaining the composition.

Chemical composition. The chemical composition of certain feldspars (ideally pure) is shown in Table 1.

TABLE 1

COMPOSITION OF FELDSPARS

	SiO_2	Al_2O_3	CaO	Na_2O	K_2O	Total	Sp. Gr.
I......	64.7	18.4	....	....	16.9	100	2.54
II......	68.7	19.5	....	11.8	....	100	2.62
III......	43.2	36.7	20.1	....	....	100	2.76
IV......	55.6	28.3	10.4	5.7	....	100	2.69

I. Orthoclase (and microcline).
II. Albite.
III. Anorthite.
IV. Labradorite ($An_{50}An_{50}$, equal molecular amounts of albite and anorthite).

The composition of the intermediate plagioclase feldspars varies with the proportions of the end members present; an example of a feldspar labradorite having equal proportions is given in IV. The other substances, such as iron oxide, etc., shown in feldspars by chemical analyses, are due to impurities.

Blowpipe and chemical characters. A fine splinter fuses before the blowpipe, with difficulty, to a globular ending—more easily if it is anorthite or one of the varieties rich in anorthite than if it is albite or orthoclase. The flame shows the persistent yellow coloration of sodium; only rarely is a rock feldspar a pure enough orthoclase to give the violet flame of potassium. In other words, the minor amount of sodium that occurs in all rock-making potassium feldspars masks out the delicate violet flame of potassium; consequently, in order to ascertain whether potassium is present the flame must be viewed through a blue glass.

Orthoclase and albite are not acted on appreciably by ordinary acids, but the plagioclase feldspars are more easily soluble in acid as their calcium content increases; thus labradorite is very slowly dissolved, whereas anorthite is slowly dissolved and yields gelatinous silica.

Alteration. Under the action of various agents the feldspars become altered, the newly formed substances being determined in part by the nature of the agents and in part by the composition of the feldspar attacked. Some of these changes and products are complex, and their nature and significance have not yet been sufficiently studied for us to understand them, but some of the simpler and more important ones are as follows.

When feldspars are acted on by water carrying dissolved carbon dioxide, as by surface waters percolating downward through the rocks or by hot waters rising from the depths below, they may be altered to kaolinite or muscovite, accompanied by separation of free silica and alkaline carbonates. These changes can be expressed chemically as follows:

$$\overset{\text{Orthoclase}}{2KAlSi_3O_8} + \overset{\text{Water}}{2H_2O} + \overset{\substack{\text{Carbon}\\\text{dioxide}}}{CO_2} = \overset{\text{Kaolinite}}{H_4Al_2Si_2O_9} + \overset{\text{Quartz}}{4SiO_2} + \overset{\substack{\text{Potassium}\\\text{carbonate}}}{K_2CO_3}$$

$$\overset{\text{Orthoclase}}{3KAlSi_3O_8} + \overset{\text{Water}}{H_2O} + \overset{\substack{\text{Carbon}\\\text{dioxide}}}{CO_2} = \overset{\text{Muscovite}}{H_2KAl_3(SiO_4)_3} + \overset{\text{Quartz}}{6SiO_2} + \overset{\substack{\text{Potassium}\\\text{carbonate}}}{K_2CO_3}$$

What determines whether the potassium will be completely removed from the feldspar, so that kaolinite is formed, or only partly, so that muscovite is the resultant product, is not clearly understood. In a general way, one can say that weathering induced by surface waters generally forms kaolinite, but under the special conditions of lateritic weathering (page 179) the kaolinite thus formed is altered to gibbsite, $Al(OH)_3$, or the feldspars even alter directly to gibbsite.

The change to muscovite is more likely to be a deep-seated affair and is especially effective in processes of metamorphism. In mines it is often seen that the solutions that deposited the ores have altered the rocks inclosing the ore bodies, in some places to kaolinite, in others to a minutely fibrous form of muscovite (sericite), and in others again to other products. Mainly owing to this process the rocks adjacent to ore bodies have generally become changed from their original fresh condition to a mass of alteration products.

All feldspars undergo changes similar to those mentioned, but the feldspars that contain calcium alter more complexly, since calcite may also be formed.

According as the change to muscovite or kaolinite is more or less complete, the feldspars lose their original vitreous luster and become dull and earthy; if the change is pronounced, they are soft and can be cut or scratched with the knife or even with the fingernail. As a result of certain changes, the plagioclase feldspars take on a faint, glimmering luster, become semi-translucent and of a pale bluish or grayish tone, lose to a great extent their property of cleavage, and resemble wax or paraffin. These changes as a rule do not take place uniformly throughout the entire crystal; in some the border is altered, in some only the center is attacked, and in others, especially the plagioclase feldspars, zones between the two are altered. If the feldspars of a rock do not show bright, glistening cleavage surfaces, they are manifestly in a more or less altered state. These alterations of the feldspars are of great importance in many geologic processes, including the formation of soils.

Besides these alterations, others also are known. Under some circumstances the feldspars are changed into zeolites, and in metamorphic processes the feldspars containing calcium may react with other minerals to form epidote and garnet, changes that are mentioned elsewhere (page 190).

Occurrence. The feldspars are widely distributed, more so than any other minerals. They occur in all classes of rocks: in most igneous rocks, such as granites, syenites, porphyries, and felsite lavas; in sedimentary rocks in certain kinds of sandstones and conglomerates; and in metamorphic rocks in gneisses. So far as is now known, the crust of the Earth, below the veneer of sedimentary beds, is composed chiefly of granites and gneisses, in which feldspars preponderate. It is no exaggeration to say that there is more feldspar in the world than any other substance.

Determination. In general, the two cleavages nearly at right angles, the vitreous luster, the light color, and the hardness, which resists the point of the knife, enable one to recognize the feldspar grains in rocks and to distinguish them from other common minerals, especially quartz, with which they are usually associated. Sometimes the crystal form is of assistance, especially in porphyries. In addition, one or more of the chemical and physical properties enumerated above can be determined on separated fragments, if the feldspar grains are sufficiently large.

Determining the different varieties of feldspar that are present in a rock is a difficult task, however, when only megascopic means are used. The remarks made under the heading of color are sometimes of assistance. If the cleavage surfaces are closely examined with a lens and the delicate striae of albite twinning can be seen, then we know that plagioclase feldspar is present, since orthoclase does not have this twinning, as previously explained. The only practical exception to this rule is that the large, or even huge, crystals of potassium feldspar in granite pegmatite dikes are generally not orthoclase but microcline, the triclinic form of potassium feldspar, and a good cleavage surface of microcline examined in a strong light with a powerful lens may show a minute, scarcely perceptible, multiple twinning like the albite twinning.

If multiple twinning cannot be detected on the cleavage surface of a feldspar, we cannot therefore safely conclude that the feldspar is orthoclase and not plagioclase, because the albite twinning lamellae, as already stated, are commonly so thin that they cannot be seen with the lens; or they may even be wanting. As the grain of rocks grows finer, the twinning striae become increasingly difficult to detect. However, a good training of the eye by studying rocks in which striae are easily seen on the feldspars helps greatly and eventually enables one to perceive the striae clearly where at first they could not be seen. The tendency on the part of geologists to relegate the determination of rocks to the microscopic examination of thin sections has led to a great neglect in training the eye to be able to determine minerals in rocks megascopically, and to a corresponding loss of efficiency in the field.

If albite twinning is clearly perceptible in several feldspar grains of a rock, we can safely conclude that much plagioclase is present; plagioclase may, indeed, be the only feldspar present. If albite twinning cannot be seen, plagioclase may or may not be present.

Other means that can be used are the determination of the specific gravity, the behavior before the blowpipe, and the reaction with acids, as previously mentioned, and the chemical tests for sodium, potassium, and calcium, which suggest themselves to those experienced in analytical chemistry. Further information on the subject should be sought in the special manuals of determinative mineralogy.

The Feldspathoid Group

The feldspathoid group is so named because, like the feldspars, it consists of minerals that are silicates of aluminum with sodium, potassium, and calcium and that accompany or take the place of the feldspars in the making of rocks. Unlike feldspars, the feldspathoids are relatively rare, being restricted to special kinds of igneous rocks so rare that they make up less than one percent of the total of the igneous rocks of the world. Thus feldspathoids are far less important than the feldspars, but in problems regarding the origin of igneous rocks, they are highly significant. The more abundant feldspathoids are *nepheline* and *leucite*; less common are *sodalite, noselite,* and *hauynite.*

Nepheline. Nepheline crystallizes in short, thick, hexagonal prisms having a flat base and top, but it rarely shows distinct crystal form in rocks. It generally occurs as irregular grains, like quartz. Its normal *color* is white, but it is generally gray, ranging from light smoky to dark in tone; some is flesh colored or brick red. The white color may shade into yellowish, the gray into bluish or greenish. The *streak* is light—not characteristic. Nepheline is translucent; its *luster,* when fresh, is vitreous, with a somewhat oily or greasy cast. Larger grains have weak *cleavage*; its fracture is conchoidal. Small glassy grains are therefore likely to resemble quartz. It is brittle. The *hardness* is nearly that of feldspar, 6. The *specific gravity* is 2.55 to 2.61. Its *composition* is chiefly $NaAlSiO_4$, with a small, varying amount of potassium proxying for sodium. Before the blowpipe a fine splinter fuses readily to a globule, tinging the flame deep yellow. Nepheline is readily soluble in dilute acid with formation of gelatinous silica.

On weathered surfaces of rocks, nepheline tends to be etched out, a feature that at once distinguishes it from feldspar and quartz.

Nepheline and quartz are mutually incompatible, that is, the presence of one in a rock excludes the other. This incompatibility is indicated in the following equation.

$$\underset{\text{Nepheline}}{NaAlSiO_4} + \underset{\text{Quartz}}{2SiO_2} = \underset{\text{Albite}}{NaAlSi_3O_8}$$

Sodalite. Sodalite crystallizes as isometric dodecahedrons, the form so commonly seen in garnet, but crystals are rare in rocks, as the mineral commonly occurs as shapeless grains. The color may be white, pink, or greenish gray but most commonly is blue of some shade, generally a bright sky blue to dark rich blue. Sodalite is generally translucent. *Cleavage* is dodecahedral but is not striking as a megascopic property; *fracture* is uneven to poorly conchoidal. The *luster* is vitreous to greasy; the streak is white. The *hardness* is nearly that of feldspar, 5.5 to 6. The *specific gravity* is 2.15 to 2.30. Its *composition* is $Na_4(AlCl)Al_2(SiO_4)_3$, which can be expressed also as $3NaAlSiO_4 \cdot NaCl$, but this formula does not mean that sodalite is a mixture of nepheline and common salt molecules; it is a definite chemical compound into which the chlorine enters. Sodalite fuses rather easily before the blowpipe, with bubbling, and colors the flame yellow. It is easily soluble in dilute acids with formation of gelatinous silica; in nitric acid solution it gives with silver nitrate the test for chlorine.

A notable property of sodalite is its fluorescence in ultraviolet light, and this property has ingeniously been used to determine the field distribution of sodalite in the nepheline syenite mass of Red Hill, New Hampshire.

Leucite. Leucite is a feldspathoid crystallizing in isometric trapezohedrons, a form illustrated in garnet; the crystals, when imperfect, appear spherical. It has imperfect *cleavage*; conchoidal *fracture*; white to gray *color*; and vitreous *luster*. *Hardness* is 5.5 to 6, and *specific gravity* is 2.5. Before the blowpipe it is infusible; when mixed with powdered gypsum it gives the flame the violet color of potassium. It dissolves in acids without gelatinizing. Its *composition* is $KAl(SiO_3)_2$. It occurs almost wholly in lavas and is nowhere common except in those of central Italy and Java, where the magmas are characterized by a high content of potash. The most noted occurrence is in the lavas of Vesuvius, in some of which leucite occurs abundantly in good-sized, well-shaped crystals of the form illustrated in Fig. 29 under garnet. Large crystals altered to other minerals, however, occur in certain syenites and related rocks in Arkansas, Montana, Brazil, and elsewhere.

Hauynite and noselite. Hauynite and noselite show characters like sodalite, but they differ from it in containing the radical SO_4 in the place of chlorine. The best method of detecting them is to test for sulfuric acid with barium chloride in their nitric acid solution. They differ from each other only in the fact that noselite is the pure sodium compound, while hauynite has part of its sodium replaced by calcium.

Alteration. The feldspathoids, like the feldspars, are readily altered by weathering as well as by other agents. They become converted into kaolinite or muscovite and also, commonly, into zeolites. Alteration into zeolite occurs very generally; hence when feldspathoids are heated in a closed glass tube they are likely to yield water.

The feldspathoids in rocks are best determined by chemical means. Except leucite, they all yield gelatinous silica and can be tested for as described later under mineral tests. Nepheline is easily confused with quartz, which it may closely resemble as seen in the rocks. However, its association with other minerals and the appearance of the rocks in which it chiefly occurs help in arousing suspicion of its presence, and this is readily confirmed by its solubility in acids. A valuable clue to its presence in rocks, as already indicated, is afforded by the fact that it is easily susceptible to weathering, and consequently the weathered surface of a nepheline-bearing rock is marked by innumerable small pits formed by the etching out of the nepheline.

Micas

The micas form a natural group of rock minerals that are characterized by a remarkably fine cleavage in one direction and by the thinness, toughness, and flexibility of the elastic plates, or laminae, into which this cleavage permits them to be split. For the practical purpose of megascopic rock study, they can be divided into two main groups: *light-colored* micas, or *muscovite* and related varieties; and *dark-colored* micas, or *biotite* and related varieties.

Form. Micas crystallize in six-sided tablets with flat bases. These tablets seem to be short hexagonal prisms (see Fig. 15); in reality, as can be shown by optical methods, they are monoclinic. Their side faces are rough and striated; the flat bases, which are usually cleavage faces, are bright and glittering. In some crystals two of the side faces are much elongated, as in Fig. 16. Although distinct crystals are often seen in rocks, particularly igneous rocks, the micas are much more common as irregular flakes and scales having flat, shining cleavage faces. Some folia, or leaves, are curled or bent.

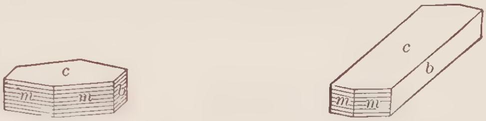

Fig. 15 Fig. 16

Biotite crystals.

Cleavage. The cleavage is perfect in a direction parallel to the base. This property, combined with the flexibility, transparency, and toughness of the mineral, makes the large crystals and sheets of muscovite found in pegmatites so useful in making stove windows, lamp chimneys, etc., where ordinary glass is easily broken. When muscovite occurs as an aggregate of minute scales its cleavage is not so apparent but can generally be seen by close observation.

Color, luster, and hardness. Muscovite is colorless, white to gray or light brown, or with greenish tones. The other light-colored micas are similar, except that the lithium mica (lepidolite) occurring in the pegmatites of some localities is generally pink or lilac-colored. In thin sheets the micas are transparent.

Biotite and its congeners are black; in thin sheets they are translucent, with strong brown, red-brown, or deep-green colors. Phlogopite is pale brown, but some is coppery. The luster of micas is splendent, on cleavage faces sometimes pearly, and in the minutely crystalline variety of muscovite called *sericite* it is silky. The hardness ranges from 2 to 3; all are easily scratched with the knife.

Chemical composition. Chemically, the rock-making micas fall into two main groups: one, exemplified by the dark-colored biotite, contains iron and magnesium; and the other, of which muscovite is the most prominent member, is devoid of these elements. The micas are complex in composition, being silicates of aluminum and an alkali element (generally potassium), and containing hydroxyl and fluorine. The two main groups can be represented as follows:

$$\text{Muscovite} = KAl_2(AlSi_3O_{10})(OH,F)_2$$

$$\text{Biotite} = K(Mg,Fe'')_3(AlSi_3O_{10})(OH,F)_2$$

The other members of the muscovite group are *paragonite,* a rare mica in which sodium replaces potassium—it is the sodium analogue of muscovite; and *lepidolite,* in which lithium partly replaces the potassium.

In the biotite group, *phlogopite* is the variety nearly free from iron and is thus a magnesian mica; its lack of iron accounts for the fact that it is lighter in color than biotite. *Lepidomelane,* on the contrary, is rich in iron, especially ferric iron. Biotite, intermediate in composition between phlogopite and lepidomelane, is by far the most abundant of the dark micas.

Table 2 gives analyses of the chemical composition of the micas.

Blowpipe and chemical characters. Most micas whiten before the blowpipe and fuse on the edges, when in thin scales. Lepidomelane fuses to a black, magnetic globule. Heated in the closed glass tube, micas yield very little water; this trait helps to distinguish them from chlorites and other micaceous rock minerals. When thin scales are treated with a little boiling concentrated sulfuric acid in a test tube, muscovite and related light-colored kinds are scarcely attacked, but biotite and its congeners are decomposed, the scales losing their luster and transparency while the acid becomes turbid.*

Lepidomelane is soluble in hydrochloric acid, depositing silica in scales. This character is important in serving to distinguish it from the other micas. Lithium micas impart a red color to the blowpipe flame, and paragonite imparts the yellow color of sodium.

Alteration. Biotite under the action of weathering becomes leached of its bases, loses its elasticity, and eventually becomes decolorized to grayish silvery flakes. In the intermediate stages it has a bronzy luster and is golden yellow; in fact, it is often mistaken by the inexperienced for flakes of gold. Muscovite strongly withstands the

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* Care should be used in making this test not to bring the hot acid in contact with water, because mixing will take place explosively.

TABLE 2

ANALYSES OF MICAS

	I	II	III	IV	V	VI
SiO_2	44.6	46.8	48.8	36.0	39.6	32.1
Al_2O_3	35.7	40.1	28.3	18.8	17.0	18.5
Fe_2O_3	1.0	....	0.3	5.6	0.3	19.5
FeO	1.0	....	0.1	14.7	0.2	14.1
MgO	....	0.6	....	9.8	26.5	1.0
CaO	0.1	1.3	0.1	0.6	....	....
Na_2O	2.4	6.4	0.7	0.4	0.6	1.5
K_2O	9.8	....	12.2	9.3	10.0	8.1
Li_2O	....	....	4.5	....	....	....
H_2O	5.5	4.8	1.7	2.5	3.0	4.6
F	0.7	....	5.0	0.3	2.2	....
XyO[1]	....	....	....	1.9	1.2	1.4
Totals	100.8	100.0	101.7	99.9	100.6	100.8

[1] XyO represents small quantities of nonessential oxides present.

I. Muscovite, Auburn, Maine.
II. Paragonite, the Alps.
III. Lepidolite, Hebron, Maine.
IV. Biotite, from granite, Yosemite, California.
V. Phlogopite, Burgess, Ontario.
VI. Lepidomelane, from nepheline syenite, Litchfield, Maine.

process of weathering, and its scales commonly occur in soils that are made up of weathered rocks, the other constituents of which are greatly altered.

Occurrence. The common micas are widely distributed as rock components. *Biotite* is abundant in many igneous rocks, especially those rich in feldspar, such as granites and syenites. In ferromagnesian rocks—basalts, gabbros, and so on—it is less common. It occurs also in many felsite lavas and porphyries. It is abundant in many metamorphic rocks, such as gneisses and schists, and is frequently one of the products of contact metamorphism exerted by igneous rocks. Because it alters easily, it does not figure as a component of sedimentary beds, but its bleached derivative is probably abundant. *Phlogopite,* the member of the biotite group containing little iron, is rare in igneous rocks; it occurs chiefly as a product of metamorphism in impure marbles and dolomites.

Muscovite occurs in granites, especially in pegmatites and in miarolitic druses of granites, and in places where the igneous rocks have been subjected to later fumarole action, which has furnished water and fluorine. It is especially common in the metamorphic rocks, being widely distributed in gneisses and schists. In some metamorphic rocks, especially in phyllites, it is in the form of aggregates of minute scales that have a silky luster and lack the ordinary megascopic characters of the mineral, such as its cleavage; this important variety is called *sericite*. When feldspars are altered to muscovite rather than to kaolinite, sericite is commonly formed. In sedimentary rocks, such as conglomerates and sandstones, muscovite is fairly common, being an unchanged remnant of the original rocks from which the material of the sedimentary rocks was derived. *Lepidolite* is practically restricted to granite pegmatites and is constantly accompanied by tourmaline. *Paragonite* is known from only a few places—in certain schists, where it plays the role ordinarily taken by muscovite.

Determination. From most rock minerals the micas can at once be distinguished by their appearance, high luster, and eminent cleavage. The last-named quality and the hardness can readily be tested in the field by the knife point. From chlorite and from talc, which resemble them, they are distinguished by the elasticity of their split-off laminae, thin flakes of the chlorites and talc being flexible but not elastic. From *chloritoid*, a micaceous-appearing mineral of a gray or green color, which is a hydrous silicate of aluminum, magnesium, and iron occurring as distinct crystals in low-rank metamorphic rocks, they can readily be distinguished because of the superior hardness (6.5) and the brittleness of chloritoid. The several varieties of mica are best discriminated by the chemical and blowpipe tests already mentioned.

PYROXENES

The pyroxene group embraces a number of important minerals that have two properties in common: (1) they are of the metasilicate type; and (2) they have closely related crystal form and a prismatic cleavage of 87° and 93°. As rock-making minerals they are extremely abundant in the igneous rocks; in fact, some igneous rocks (logically called "pyroxenites") are composed almost entirely of pyroxene.

It is often difficult to recognize pyroxene in the rocks and distinguish it from several other minerals purely by megascopic methods, and it is generally impossible to tell apart, by such means, the many varieties recognized by mineralogists and petrographers. These varieties differ chiefly in chemical composition and optical properties, and the distinctions can be determined only by chemical and optical methods.

The pyroxenes can be divided for megascopic purposes into the following subgroups, dependent on color, behavior before the blowpipe, and chemical reaction for calcium, as described later: *hypersthene, diopside, augite,* and *aegirine.*

Form. Hypersthene crystallizes in the orthorhombic system, the other pyroxenes in the monoclinic system; however, this distinction is not practically useful in megascopic work, as hypersthene is rarely well enough crystallized to enable one to determine its system. The common form in which the monoclinic rock pyroxenes crystallize is a prism, short and thick as a rule, though some are longer and more slender. Such a prism is shown in Fig. 17, the ends modified by pyramidal faces. Generally, however, the edges of the prism *mm* are truncated by a front pinacoid *a* and a side pinacoid *b*; these truncations may be small, with the result that *a* and *b* are narrow faces (Fig. 18), but as a rule the pinacoids are broad and prism faces narrow. Although these faces are generally well developed and lustrous, the pyramidal faces may be imperfect or wanting, the crystal being rounded at the ends; rarely, other pyramidal faces are present and the ends are more complex than shown in the figures. The augite crystals in igneous rocks, especially in porphyries and lavas, generally have the appearance and development shown in Fig. 19. The most important crystallographic feature of the pyroxenes is that the angle *m* on *m* is nearly a right angle, 87°, so that the prism is nearly square in cross section, or, when truncated by *a* and *b*, eight sided, as shown in Fig. 20. Besides occurring as prismatic crystals, the pyroxenes are common as grains or as more or less irregularly shaped masses; this is usually the way they appear in massive igneous rocks, such as gabbros and peridotites.

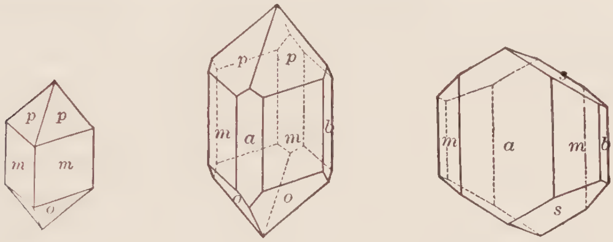

Fig. 17 Fig. 18 Fig. 19

Pyroxene crystals.

Cleavage and fracture. As previously mentioned, the pyroxenes have a cleavage parallel to the prismatic faces, nearly at right angles, as shown in Fig. 21. This feature is fundamental and serves to distinguish the pyroxenes from the amphiboles. This cleavage is generally good but not perfect. Some varieties as the result of pressure to which they had been subjected have a good parting, resembling cleavage, but its direction differs from that of the cleavage. The parting causes the mineral to appear lamellar, almost micaceous, as in the pyroxenes of some gabbros. Fracture is uneven; the mineral is brittle.

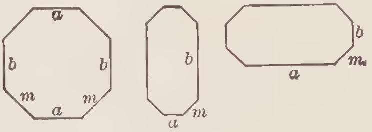

FIG. 20. Cross sections of pyroxene.

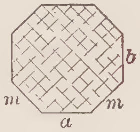

FIG. 21. Cross section of pyroxene, showing position of cleavage.

Color and luster. The pyroxenes range from white through shades of green to black, according to the amount and proportion of ferrous and ferric iron present. Diopside is white when pure, rarely colorless and transparent, but is generally pale green, and more or less translucent. Augite is dull green to black. Aegirine is black and opaque. The luster, which may be wanting, is vitreous. The streak is white to gray green.

Hardness and specific gravity. The hardness ranges from 5 to 6. Some varieties can just be scratched by the knife. The specific gravity ranges, chiefly as the result of the iron present, from 3.2 to 3.6.

Chemical composition. Pyroxenes are solid solutions composed of the metasilicate molecules $MgSiO_3$, $Fe''SiO_3$, $CaMg(SiO_3)_2$, $CaFe''(SiO_3)_2$, $NaFe'''(SiO_3)_2$, and $R_2'''O_3$, in the last of which $R = Al$ and Fe'''. These molecules are capable of forming several solid-solution series, all of which have essentially the same crystal form and many similar physical properties. The *hypersthene* group, comprising the *orthopyroxenes* (so called because they are orthorhombic), are solid solutions of $MgSiO_3$ and $FeSiO_3$; their formula is $(Mg, Fe)SiO_3$.

All the other pyroxenes are termed *clinopyroxenes*, because they crystallize in the monoclinic system. The simplest clinopyroxene is *diopside*, $CaMg(SiO_3)_2$. *Augite* is a complex solid solution consisting commonly of about 60 per cent of the diopside molecule and 40 per cent of the hedenbergite [$CaFe(SiO_3)_2$] molecule, Al_2O_3, Fe_2O_3, and other molecules.

Blowpipe and chemical characters. Hypersthene varies in fusibility: if it contains little iron (this variety is called *enstatite*), it is almost infusible in the blowpipe flame; it is fusible with difficulty if it contains much iron, becoming black and slightly magnetic. The other pyroxenes are much more fusible (4) and melt quietly or with little intumescence to glassy globules, the color of which depends on the amount of iron: diopside is nearly colorless; augite and aegirine are black; they are also magnetic. Aegirine fuses quietly, coloring the flame yellow. The pyroxenes are but slightly acted on by acids, those containing iron more so than those without.

The differences in the chemical composition are shown in Table 3.

TABLE 3

ANALYSES OF PYROXENES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	XyO[1]	Total
I...	53.1	1.0	....	17.8	24.8	2.7	....	0.4	99.8
II...	58.0	1.3	....	3.1	36.9	....	....	0.8	100.1
III...	55.1	0.4	....	1.1	18.1	25.0	0.4	0.2	100.3
IV...	51.1	2.0	1.3	12.3	10.0	22.1	....	0.4	99.2
V...	47.0	9.8	4.5	4.1	16.0	19.0	....	....	100.4
VI...	51.4	1.8	23.3	9.4	0.3	2.0	11.9	0.1	100.2

[1] XyO represents small quantities of other oxides.

I. Hypersthene, Romsaas, Norway.
II. Hypersthene (var. enstatite), Bamle, Norway.
III. Diopside, DeKalb, New York.
IV. Common pyroxene, Edenville, New York.
V. Black augite, Vesuvius lava.
VI. Aegirine, from syenite, Hot Springs, Arkansas.

Alteration. The pyroxenes are prone to alter into other substances, the nature of which depends partly on the kind of process to which they are subjected and partly on their own composition. Thus under the action of weathering they may be converted into carbonates, such as calcite. Pyroxenes containing much iron may break down completely into hydrated iron oxides, such as limonite, and carbonates.

Another important change is one that they undergo when subjected to metamorphic processes. In this change they are transformed to masses of fibrous or felty hornblende needles and prisms, usually of distinct but variable green colors. This process is of great geologic importance, for by means of it whole masses of pyroxenic rocks, generally of igneous origin, such as basalts, gabbros, and peridotites, have been changed into amphibolic rocks, to which various names, such as greenstone, greenstone schist, and amphibole schist, have been applied. The process is further mentioned under metamorphism, and under gabbro, dolerite, greenstone, and amphibolite.

Occurrence. The pyroxenes occur chiefly in igneous rocks, especially in those rich in lime, iron, and magnesia. They should always be looked for in the dark-colored rocks of this class. They are sparse in most igneous rocks that contain much quartz; hence they are rare in granites, felsite porphyries, and felsite lavas.

Augite is common in basaltic lavas and allied rocks as well-formed crystals; but in gabbros and peridotites it generally appears as irregularly shaped grains. Hypersthene is prominent in some varieties of gabbro and peridotite. Aegirine occurs chiefly in nepheline syenites and the variety of felsite lava called phonolite. Some normal syenites and related rocks contain diopside-like pyroxene.

In the metamorphic rocks, diopside, white or pale greenish and transparent, occurs in recrystallized impure limestones and dolomites, either as scattered well-formed crystals or aggregated into large masses. Being readily decomposed by weathering, pyroxenes play no part in sedimentary beds.

Determination. If the mineral under examination is in well-formed crystals, careful observation usually shows whether or not it is a pyroxene, by the presence or absence of the forms previously described. Whether the cross section of the prism is square or not should especially be noted. Common minerals in rocks with which pyroxenes may be confused are hornblende, epidote, and tourmaline. The lack of good cleavage, the superior hardness, the high luster, coal-black color, and triangular shape of the prism cross section of tourmaline readily distinguish that mineral from pyroxene. Epidote has one perfect cleavage and one poor one; it is much harder, 6 to 7; although green, it commonly has a yellow tone, giving a yellowish green; before the blowpipe it intumesces when fusing. The distinction between pyroxene and hornblende is more difficult to recognize and is treated in the section devoted to hornblende.

To distinguish the several varieties of pyroxene from one another, the blowpipe tests should be used in conjunction with the color of the mineral. The hypersthenes are distinguished from other pyroxenes by a chemical test which proves the absence of calcium or its presence in only minute quantity. This test is best done by making a small fusion with soda as described in the chapter treating of mineral tests.

The most expeditious and certain way of distinguishing the several pyroxenes, however, is the immersion method, in which crushed particles of the minerals are placed in oils of known refractive indices and are examined under the petrographic microscope.

AMPHIBOLES

The amphiboles are a group of silicate minerals that, like the pyroxenes, are of the metasilicate type. All amphiboles have a characteristic crystal form and a prismatic cleavage of 55°; they are closely related in many other physical properties. As in the pyroxene group, many varieties are recognized, dependent on differences in chemical composition and physical properties, especially optical ones, which are impossible to distinguish by the unaided eye and many of which cannot be distinguished by ordinary megascopic tests.

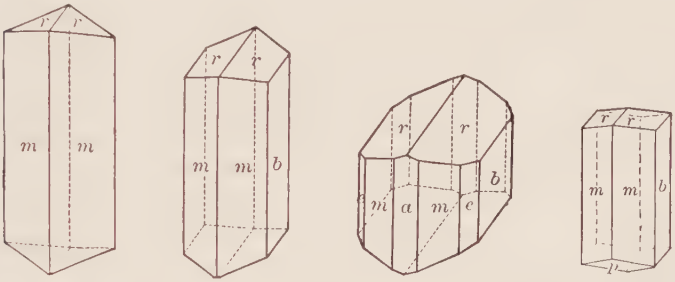

FIG. 22 FIG. 23 FIG. 24 FIG. 25

Amphibole crystals.

For practical work in megascopic petrography the following amphiboles can be recognized: *tremolite, actinolite, common hornblende,* and *arfvedsonite*. These are distinguishable by their colors, associations, and behavior before the blowpipe.

Form. Amphiboles crystallize in the monoclinic system. The crystals are generally long and bladed, formed by two prisms *mm* that meet at angles of 55° and 125°. Some have terminal faces *rr,* as in Fig. 22, but many crystals are imperfect at the ends and have no terminal faces; such imperfect crystals are common in rocks. The side pinacoid *b* is generally present, truncating the prism edge, and the crystal has a nearly hexagonal cross section, as in Fig. 23. Rarely, the front pinacoid *a* is present, as in Fig. 24. The hornblendes conspicuous as phenocrysts in andesites commonly have a fairly long prism and appear as in Fig. 25; they are the hornblendes that as a rule have distinct terminal planes. If present, the prismatic faces *mm* and the pinacoid *b* tend to be shining; the ends are generally dull.

Amphibole rarely occurs in rocks as crystals whose planes can be distinctly seen. If recognizable crystals occur, they are mostly the hornblendes forming the phenocrysts in lavas and the amphiboles occurring in limestones and dolomites altered by metamorphism. Amphibole commonly appears as long, slender blades with irregular, rough ends, typically in the hornblende schists where the crystals are aggregated together in more or less parallel position. They may be so small as to be mere shining needles so minute that the individual prisms can hardly be discerned with the lens; the aggregate then has a silky appearance. In felsitic lavas and porphyries the hornblende phenocrysts range from short prisms, like those in the figures, to slender needles. In massive granular rocks, such as diorite, the amphibole generally occurs as irregular grains.

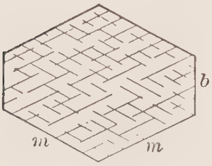

Fig. 26. Cross section of amphibole, showing position of cleavage.

Cleavage. Amphiboles have a perfect cleavage parallel to the prism faces *mm*, as illustrated in the cross section, Fig. 26. Like the prism faces *mm*, these cleavages meet at angles of 125° and 55°, a feature of great importance in distinguishing all the amphiboles. The glittering prismatic faces seen on the blades and needles of broken rock surfaces are commonly due to this perfect cleavage. The fracture is uneven.

Color and luster. The color varies with the amount of iron from purest white and gray in tremolite through gray green or bright green in actinolite to dark green and black in common hornblende. Arfvedsonite is black. Some varieties of amphibole in igneous rocks seem to be black but are really deep brown. The amphiboles vary from opaque in the deeper-colored varieties to translucent in the lighter ones. The luster is bright and vitreous to somewhat pearly on the cleavage surfaces; in very fine, needle-like or fibrous varieties it is silky. Some of the black amphiboles have almost a metallic luster. The streak of all amphiboles, regardless of their color, is white to gray green or brownish.

Hardness and specific gravity. The hardness ranges from 5 to 6; some specimens can be scratched with the knife. The specific gravity ranges, chiefly with the amount of iron, from 3.0 to 3.5.

Chemical composition. Amphiboles, like pyroxenes, can be considered to be metasilicates, salts of H_2SiO_3 in which the hydrogen atoms are replaced by calcium, magnesium, iron, sodium, and radicals in which aluminum plays a prominent part and which contain hydroxyl (OH) and fluorine. When calcium is present, it replaces one fourth of the hydrogen atoms. Thus, although amphiboles resemble pyroxenes in being metasilicates and composed of the same elements, they differ from them in being much more complex and in containing hydroxyl and fluorine. Their compositions, as a rule, are too complex to be represented by simple formulas.

Thus, *tremolite* is represented by $H_2Ca_2Mg_5(SiO_3)_8$, and *actinolite* by $H_2Ca_2(Mg, Fe'')_5(SiO_3)_8$, in which ferrous iron "proxies" for magnesium.

Common hornblende, or *hornblende* for short, differs from actinolite in containing aluminum or ferric iron, generally both. *Arfvedsonite*, which contains chiefly sodium, calcium, and ferrous iron, corresponds among the amphiboles to aegirine, the sodium-iron pyroxene, although a rare variety, *riebeckite*, corresponds more nearly in composition.

Glaucophane is a rare variety of amphibole, distinguished from all the others by its deep-blue color. It occurs only in a rare variety of amphibole schist called glaucophane schist, which is described under amphibolites.

The chemical composition of the amphiboles is illustrated in Table 4.

TABLE 4

ANALYSES OF AMPHIBOLES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	H_2O	F_2	XyO[1]	Total	Sp. Gr.
I...	57.5	1.3	0.2	0.2	24.9	12.8	0.7	1.3	0.8	0.6	100.3	....
II...	56.1	1.2	0.8	5.5	21.2	12.1	0.2	1.9	0.1	0.6	99.7	....
III...	43.1	11.4	4.7	11.4	12.4	11.6	2.0	1.6	....	2.1	100.3	3.268
IV...	43.8	4.4	3.8	33.4	0.8	4.6	8.1	0.1	....	1.5	100.5	....
V...	57.7	12.0	1.2	5.4	13.0	1.0	7.0	2.3	....	0.7	100.3	3.085

[1] XyO represents small quantities of minor components.

I. Tremolite, Richville, near Gouverneur, New York.
II. Actinolite, Greiner, Tyrol.
III. Hornblende from the type syenite, Dresden, Germany.
IV. Arfvedsonite, Kangerdluarsuk, Greenland.
V. Glaucophane, Zermatt, Switzerland.

Blowpipe and chemical characters. Tremolite, actinolite, and common hornblende melt quietly or with little intumescence before the blowpipe, fusing rather easily at 4. The color of the bead depends on the amount of iron: tremolite gives a bead nearly colorless; actinolite, green or brown; common hornblende, dark and shining. Some common hornblende colors the flame yellow, indicating sodium.

Arfvedsonite fuses easily (3.5), colors the flame a strong, persistent yellow, intumesces decidedly (thus differing from aegirine), and yields a black, shining, magnetic bead. The amphiboles are only slightly attacked by ordinary acids, those richer in iron being the most easily attacked.

Alteration. The amphiboles alter similarly to the pyroxenes. Under the action of various agents they are changed into serpentine or chlorite or both, with an accompanying formation of carbonates, sometimes of epidote, and also of quartz. Under the continued action of weathering they break down into limonite, carbonates, and quartz. Thus on much-weathered rock surfaces, only rusty-looking holes and spots may be left to indicate the former presence of amphiboles.

Occurrence. Amphiboles are common and widely distributed minerals playing an important role in igneous and metamorphic rocks. The presence of water, hydroxyl, and fluorine in the amphiboles shows that they are not formed as simply as the pyroxenes but require the presence of mineralizing vapors. Thus they cannot be artificially formed by allowing simple dry fusions containing their constituent silica and metallic oxides to cool and crystallize; pyroxenes are produced instead. Moreover, if hornblendes are fused and the melt is allowed to crystallize, pyroxenes and magnetite are obtained in their place, because the necessary water and fluorine have escaped. Evidently, pressure is required to keep the fluorine and the gaseous water on the job.

Tremolite occurs chiefly in the impure crystalline limestones and dolomites in contact-metamorphic zones and in schistose metamorphic rocks. In such occurrences it may have an extraordinarily finely fibrous structure and is capable of being split into long, flexible fibers of great fineness and strength, forming what is known as *amphibole asbestos.* Actinolite and other amphiboles are found in places in the asbestos form. Most of the asbestos of commerce, however, is a fibrous variety of serpentine.

Actinolite has its true home in the schists; it is the characteristic light green to bright green amphibole of many amphibole schists and greenstones: in many of these it is secondary after the original pyroxene of former gabbro and trap rocks, as described under uralite.

Common *hornblende* occurs both in igneous and in metamorphic rocks. It is found in granites, common syenites, diorite, and some varieties of peridotite. It is abundant as phenocrysts of felsitic intrusive porphyries and lavas. In dark traps and basalt lavas it is rare. In the metamorphic rocks it occurs in gneisses and is the prominent mineral of the hornblende schists.

Arfvedsonite occurs in nepheline syenites and in rare porphyries.

Uralite is a fibrous or fine-needle-like, columnar hornblende, secondary after pyroxene, and, as mentioned under that mineral, produced from it by metamorphic processes. The outward crystal form of the pyroxene is retained, but the substance is now hornblende in parallel bundles of needle-like prisms. Generally the secondary amphibole is in aggregates, which may be very fine and feltlike, lying in the plane of schistosity. It is especially likely to occur when basic, pyroxenic igneous rocks have been subjected to deformation in the Earth's crust attended with squeezing and shearing. It varies in composition from actinolite to common hornblende, depending on the kind of pyroxene from which it was derived. Obviously this change can not be a simple rearrangement of the pyroxene molecule, which has twice as much calcium as the hornblende and lacks the necessary water and fluorine. Calcium separates out during the alteration of pyroxene to hornblende and forms a carbonate (calcite) or other calcium mineral, and the presence of water is necessary to produce the amphibole.

In this connection the reader should consult what is said under metamorphism and the hornblende schists.

Determination. Amphibole is likely to be confused in megascopic work with pyroxene, tourmaline, and epidote. To distinguish it from tourmaline and epidote, use should be made of the various physical properties mentioned under the determination of pyroxene. The excellent cleavage distinguishes at once the amphibole from tourmaline. To distinguish amphibole from pyroxene is much more difficult, because these minerals have rather similar chemical compositions and physical properties. The following points will be found of service in this connection. If the mineral is in tolerably distinct crystals, its form should be carefully studied, especially the outline of the cross section of the prism, which can often be seen on a fractured surface of the rock, and comparison should be made with Figs. 21 and 26.

If the crystal form is imperfect or wanting, the angle at which the cleavage surfaces meet should be carefully studied, as this is a fundamental character, the cleavage prism, as already described, being nearly square in pyroxene and highly oblique in amphibole. Further, the perfection of the cleavage in amphibole and the resulting bright, glittering surfaces give indications not commonly seen in pyroxene, the cleavage of which is only fairly good. Amphibole tends also to occur in needles or long-bladed prisms; pyroxene is commonly in short prismoids or grains.

Before the blowpipe, amphibole, because of its combined water (hydroxyl), is more likely to intumesce than pyroxene (distinction of arfvedsonite from aegirine), but this cannot be relied on as a general definite test. If fluorine is obtained by a qualitative test, this is also indicative of amphibole, but many amphiboles do not contain fluorine; and the test is not one that is ordinarily in one's power to make. Finally, it is often impossible, especially in examining fine-grained igneous rocks, to tell by purely megascopic means whether the dark ferromagnesian mineral present is hornblende or pyroxene or, as often happens, a mixture of both. To recognize this difficulty frankly, it has been proposed to call the dark mineral *pyrobole* (from *pyroxene* and *amphibole*), implying thereby that one or the other is present, or both, but that they can not be certainly told apart. Examined under the microscope in a thin section or as powder immersed in an index oil, they can of course be easily and rapidly identified.

OLIVINE

Form. Olivine crystallizes in the orthorhombic system; the crystals are rather complex, as illustrated by a common form shown in Fig. 27. The form is not, however, a matter of practical importance, because olivine is rare as well developed crystals in rocks but occurs as grains or small, irregular masses composed of grains.

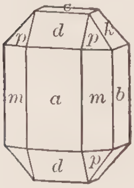

FIG. 27. Olivine.

General properties. The cleavage parallel to the side pinacoid b is poor and is not perceptible megascopically. The *fracture* is conchoidal. The *color* is green, ranging from olive green to yellow green; bottle green is common. A beautiful golden iridescence on fractured surfaces is often seen. Olivine is transparent, or translucent, but becomes brown or dark red on oxidation of its iron content and more or less opaque. It has a vitreous *luster*, white to yellowish *streak*, and a *hardness* of 6.5 to 7.0. The *specific gravity* ranges with the iron content from 3.3 to 3.5.

Chemical composition. Olivine is a group name given to a continuous series of solid solutions ranging from magnesium orthosilicate (Mg_2SiO_4) to ferrous orthosilicate (Fe_2SiO_4). The pure magnesium compound is called *forsterite*, the pure iron compound *fayalite*; these occur in rocks but are rare. The most common member is olivine, from which the group takes its name. It is abundant in basalts, gabbros, and peridotites; in all of these the olivine is strongly magnesian, being composed predominantly of the forsterite molecule. The variation in composition in the olivine group is shown in Table 5.

Blowpipe and chemical characters. Before the blowpipe, olivine is nearly infusible. Varieties rich in iron fuse and yield magnetic globules; these varieties are likely to turn red on heating. Powdered

TABLE 5

Analyses of Olivines

	SiO_2	MgO	FeO	XyO[1]	Total
I......	41.8	56.2	1.1	0.7	99.8
II......	39.9	49.2	10.5	....	99.6
III......	37.2	39.7	22.5	....	99.4
IV......	41.9	28.5	29.2	....	99.6
V......	33.6	16.7	44.4	5.0	99.7
VI......	30.1	....	68.2	1.5	99.8

[1] XyO represents quantities of other oxides, chiefly MnO.

I. Forsterite, Monte Somma, Italy.
II. Olivine, Mount Vesuvius, Italy.
III. Olivine, Montarville, Canada.
IV. Olivine, Hochbohl, Germany.
V. Hortonolite, Monroe, Orange Co., New York.
VI. Fayalite, Rockport, Massachusetts.

olivine dissolves in hydrochloric or nitric acid, yielding gelatinous silica on evaporation. The solution may be tested for iron and magnesium as directed under mineral tests.

Alteration. When olivine alters by oxidation of the iron, it turns reddish or brownish, and eventually becomes a mass of limonite, accompanied by carbonates and some form of silica. The rusty iron product is the most noticeable result of the process. Under some circumstances olivine weathers to a golden-yellowish micaceous mineral called bowlingite.

Much olivine is found to be altered into serpentine, not as the result of weathering at the Earth's surface, but by more deep-seated processes. It is more fully discussed under the head of serpentine. Other substances, such as carbonate of magnesium, magnetite, and free silica are likely to occur as by-products in the process. Other kinds of alteration of olivine are known but are of less importance.

Occurrence. Olivine is highly characteristic of the igneous rocks rich in ferromagnesian minerals. It so rarely occurs in those composed chiefly of alkali feldspars—in granites, feldspathic porphyries, and felsites—that for practical purposes it need not be sought in them. Anorthosite is the only feldspathic rock in which it may be abundant. Thus its true home is in gabbros, peridotites, and basaltic lavas. It forms also the igneous rock known as dunite, which consists almost wholly of olivine. In the basalts it usually occurs in bottle-green grains; in the gabbros and peridotites it is sometimes darkened by inclusions. Fine transparent crystals of olivine from basaltic lavas are cut for gems, commonly called peridotes. The mineral is found also in meteorites.

Forsterite (the magnesian olivine) occurs in metamorphic rocks, in crystalline limestones of dolomitic character, and in other rocks found in such associations, composed of varying quantities of other magnesian (and calcic) silicates, such as amphibole, pyroxene, and talc. Its origin is ascribable to a reaction between dolomite and quartz or silica-bearing solutions.

$$2CaMg(CO_3)_2 + SiO_2 = Mg_2SiO_4 + 2CaCO_3 + CO_2$$

Determination. The appearance, associations, and characters described above generally suffice to identify olivine readily. It can be confused with greenish, more or less transparent grains of pyroxene, but its lack of cleavage, superior hardness, and easy gelatinization in acid enable one to distinguish it from that mineral.

GARNET

Form. Garnets crystallize in the isometric system in the simple form of the rhombic dodecahedron shown in Fig. 28 or in the trapezohedron shown in Fig. 29. Some show these forms well developed and are excellent crystals; some are complicated by bevelings of the edges of the dodecahedron; commonly, however, the faces are not well developed and the garnet appears as a spherical mass or grain.

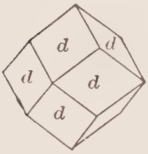

Fig. 28. Dodecahedral garnet.

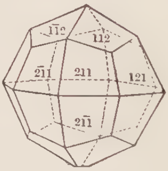

Fig. 29. Trapezohedral garnet.

Cleavage and fracture. Cleavage is absent; in some garnets occurring in sheared rocks, a parting can be seen that causes a suggestion of lamellar structure. The fracture is uneven. The mineral is brittle, but rocks composed largely of massive garnet are very tough.

Hardness and specific gravity. The hardness ranges from 6.5 to 7.5; the specific gravity, from 3.53 in grossularite to 4.32 in almandite.

Color, luster, and streak. *Grossularite* is sometimes white but is generally pale green or yellow; some is yellowish or reddish brown to brown. *Pyrope* is deep red to black. *Almandite* and most common garnet is deep red to brownish red; *andradite* is honey yellow to black, but may resemble grossularite. Unfortunately, color is not a reliable guide to the composition of the garnets. The streak is light colored and not significant. The luster is vitreous, but some garnets incline to be resinous. The light-colored garnets are transparent to translucent, the darker ones translucent or opaque.

Chemical composition. Garnets are orthosilicates of the general formula $R_3''R_2'''(SiO_4)_3$, in which the radical R'' is generally calcium, magnesium, ferrous iron, or manganese (bivalent metals), and R''' may be aluminum, ferric iron, or chromium (trivalent elements). There is therefore ample opportunity for many isomorphous combinations. The garnets most important as rock minerals are *grossularite,* $Ca_3Al_2(SiO_4)_3$, *pyrope,* $Mg_3Al_2(SiO_4)_3$, *almandite,* $Fe_3''Al_2(SiO_4)_3$, and *andradite,* $Ca_3Fe_2'''(SiO_4)_3$. These compounds, however, rarely, if ever, occur pure. Generally variable amounts of the other molecules are present in solid solution, and the resultant garnet is named after the predominating member. The particular composition assumed by the garnet is determined by the genetic conditions under which it formed. *Common garnet* is chiefly *almandite* (or almandine) admixed with more or less of the other garnet molecules.

The chemical composition of the ideal garnets and their specific gravities are given in Table 6.

TABLE 6

CHEMICAL COMPOSITION OF GARNETS

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Sp. Gr.
Grossularite	40.0	22.7	....	....	....	37.3	3.582
Almandite	36.2	20.5	....	43.3	....	....	4.325
Pyrope	44.8	25.4	....	....	29.8	....	3.554
Andradite	35.5	....	31.5	....	....	33.0	3.838

Blowpipe and chemical characters. The garnets fuse readily before the blowpipe, and in the reducing flame those containing much iron become magnetic. After fusion and grinding of the bead to powder, they dissolve in hydrochloric acid and gelatinize on boiling. Most of them are slightly attacked by acids; andradite, however, is strongly attacked. They give little or no water by heating in a closed glass tube. They are decomposed by fusion with sodium carbonate.

Alteration. Garnets are resistant to weathering. Those containing iron are much the least resistant and may alter into rusty spots of limonite and other products of decomposition. The manganese garnet (spessartine) has weathered under tropical conditions to form large bodies of manganese oxide ore in India and Brazil.

Occurrence. Almandite is widely distributed as a component of metamorphic rocks. Its most striking occurrence is in mica schists, though it is found also in related rocks—in many hornblende schists and in gneisses. It is present in some granite pegmatites, rarely in granite itself, as sporadic scattered crystals. *Pyrope,* the garnet chiefly used as a jewel, is a minor component of some peridotites and the serpentines derived from them. *Grossularite* occurs especially in metamorphosed impure limestones as a result of either contact or regional metamorphism. *Andradite* occurs in enormous masses in certain ores of contact-metamorphic origin.

Determination. The crystal form, appearance, color, and hardness generally suffice to enable one to recognize the garnets, and in case of doubt the blowpipe tests furnish confirmation. Refined tests of a quantitative nature, however, are necessary if it is desired to ascertain the exact composition of a garnet.

EPIDOTE

Form. Epidote crystallizes in the monoclinic system, the simplest form being that shown in Fig. 30; its crystals, however, tend to be more complex, having many more faces. Well-developed crystals as a rule occur only in druses in seams and cavities, and the crystal form is therefore a character not generally of much use in megascopic determination. Epidote is common as bladed prisms extended in the direction of the edge *ac.* Slender needle-like forms are common and are aggregated in bundles or sheaves. Terminations of prisms are generally rounded. It occurs also as spherical and angular grains and as aggregates of such grains.

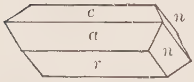

FIG. 30. Epidote.

General properties. The *cleavage* is perfect in the direction parallel to the base *c. Fracture* is uneven. Epidote is brittle with a *hardness* of 6 to 7. The *specific gravity* is 3.3 to 3.5. The *color* is green, generally a peculiar yellowish green, ranging from the distinctive pistachio green to olive and very dark green; it is rarely brownish. The *luster* is vitreous and the *streak* whitish. It is translucent to opaque.

Chemical composition. Epidote is really the name of a solid-solution series of hydrous calcium aluminum-iron silicates, whose composition is given by the formula $Ca_2(Al, Fe''')_3(SiO_4)_3(OH)$. The chemical composition is shown in the representative analyses of Table 7.

Blowpipe and chemical characters. Before the blowpipe epidote fuses easily with intumescence to a black, slaggy mass. Intense heating in the closed glass tube causes the finely powdered mineral to give off water. Epidote is only slightly acted on by hydrochloric acid, but

TABLE 7

Analyses of Epidotes

	SiO_2	Al_2O_3	Fe_2O_3	FeO	CaO	H_2O	XyO[1]	Total	Sp. Gr.
I...	37.8	27.6	7.6	0.2	24.6	1.8	0.4	100.0	3.324
II...	36.7	21.6	15.1	0.3	24.1	1.9	0.3	100.0	3.446

[1] XyO represents small amounts of other oxides.

I. Rotenkopf, Zillertal, Tyrol.
II. Bourg d'Oisans, Dauphine, France.

after fusion it dissolves and gelatinizes. It reacts with fluxes for iron and decomposes on fusion with sodium carbonate.

Occurrence. Epidote is characteristic as a product of alteration of other minerals. When basic igneous rocks undergo metamorphism of a relatively mild kind, epidote is likely to form. The most notable occurrences megascopically are those in which impure limestones containing sandy, clayey, and limonitic impurities were subjected either to general or contact metamorphism. Epidote was then developed, associated with other silicates formed at the same time, but in places it was formed so abundantly as to make extensive masses consisting almost entirely of it.

Determination. The peculiar yellow-green color, superior hardness, perfect cleavage in one direction only, and the blowpipe characters generally suffice to distinguish epidote from hornblende, pyroxene, and tourmaline, with which it might be confused. The hardness distinguishes it at once from varieties of serpentine that resemble it in color. It can be confirmed by a chemical test showing the absence of magnesium, as described in the section on mineral testing.

Zoisite. Zoisite has nearly the same chemical composition as epidote, to which it is closely related. It consists almost wholly of the calcium-aluminum molecule of epidote and contains little or no ferric iron. It is orthorhombic, but in regard to the crystals in rocks this characteristic can be ascertained only by optical methods. Zoisite occurs as aggregated blades or prisms, parallel or divergent, or as grains and masses. Its color is usually gray of varying shades. From epidotes lacking in iron it can be told only by crystallographic investigation.

Vesuvianite, or Idocrase

Vesuvianite is a tetragonal mineral whose typical habit is that of a short thick square prism, which is terminated by a pyramid cut off by a basal plane, as illustrated in Fig. 31. It occurs also in lumps or grains. The *cleavage* is poor, being perceptible, if at all, in the direction parallel to the prism faces *m*; *fracture* is uneven. The *color* is generally apple green, yellow, or brown. The *luster* is vitreous. *Hardness* is 6.5; *specific gravity*, about 3.4. The mineral is subtransparent to subtranslucent.

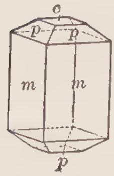

Fig. 31. Vesuvianite.

In *chemical composition* vesuvianite is a basic calcium-aluminum silicate containing hydroxyl and fluorine, but small amounts of iron and magnesium are generally present. Some varieties contain also boron. The formula is $Ca_{10}Al_4(Mg, Fe)_2Si_9O_{34}(OH, F)_4$. Before the blowpipe it fuses readily with intumescence to a greenish or brownish glass, which gelatinizes with hydrochloric acid. The fresh mineral is slightly soluble in hydrochloric acid.

As a rock-forming mineral, vesuvianite characteristically occurs in limestones that have become coarsely crystalline through the contact-metamorphic action of igneous rocks, and its formation is evidently determined by pneumatolytic emanations of steam and fluorine that issued from the magmas and reacted with the limestones. In these occurrences it is commonly associated with garnet, pyroxene, tourmaline, chondrodite, and other contact-metamorphic minerals.

Vesuvianite can easily be confused with garnet and epidote, which it resembles in color and general appearance, and with pyroxene. Its crystal form (if definitely square prismatic), its other physical characters, and its behavior before the blowpipe generally serve to distinguish it from them.

STAUROLITE

Form. Staurolite is orthorhombic and generally occurs as distinct crystals of the form shown in Fig. 32. They are stout and thick, rarely long and slender but not strikingly so. The angle of the prismatic faces *m* on *m* is 50° 40′. They are terminated by flat bases *c*, but these cannot as a rule be seen in the rock. Staurolite commonly forms cruciform twinned crystals, as shown in Figs. 33 and 34. From this striking feature its name is derived, from the Greek, *stauros*, meaning a cross.

Physical properties. Staurolite has a weak but distinct *cleavage* parallel to the face *b*; the fracture is subconchoidal. The *color* is dark reddish, yellowish brown, or almost black; the light transmitted through thin splinters appears almost blood red. The *streak* is white to gray. The *hardness* is 7 to 7.5; the *specific gravity*, 3.75.

Chemical composition. Staurolite is essentially ferrous iron aluminum silicate, $Fe''Al_4O_2(SiO_4)_2(OH)_2$. The aluminum may be partly replaced by ferric iron and the ferrous iron by magnesium, as seen in the subjoined analysis of a crystal from Franklin, North Carolina.

SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	H_2O	Total
27.91	52.92	6.87	7.80	3.28	1.59	100.37

The percentage of silica is notably low. Staurolite, in fact, is one of the rock-forming silicates that contains the least amount of silica, and this fact and its high content of alumina are significant of its place and mode of origin, namely, in metamorphosed argillaceous rocks.

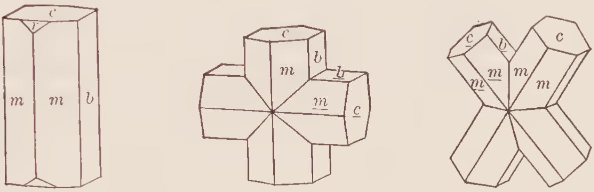

Fig. 32 Fig. 33 Fig. 34

Staurolite crystals.

Blowpipe and chemical characters. Staurolite is practically infusible before the blowpipe. It is almost insoluble in acids. It can be fused with sodium carbonate, and the resulting fusion, after being dissolved in hydrochloric acid, can be tested for aluminum, iron, and magnesium. It is easily recognized by its color, crystal form, hardness, method of twinning, and mode of occurrence.

Occurrence. Staurolite occurs in the metamorphic rocks; it is a highly characteristic mineral of the crystalline schists. It is found in phyllites of argillaceous origin and in the corresponding mica schists and gneisses. Frequently it is associated with dark-red garnets in these rocks.

ANDALUSITE

Andalusite is orthorhombic in crystallization and is generally in rough prisms, nearly square in cross section. The prisms may occur in radiated groups. The *cleavage* parallel to the prism is good; in other directions it is poor. *Fracture* is uneven to subconchoidal. The normal *color* is white to pink or red to brown, but andalusite is likely to contain impurities, especially particles of carbonaceous matter, which color it dark or even black. These are arranged in a symmetrical manner in the crystal so that, when it is broken or cut, the resulting cross section displays a remarkable symmetric pattern, such as a black cross in a white square. This feature helps to identify the mineral. It is usually subtranslucent in thin splinters. It is brittle and has a *hardness* of 7.5, a *specific gravity* of 3.2, and a whitish *streak*. The *chemical composition* is $Al_2SiO_5 = Al_2O_3 \cdot SiO_2$. It is insoluble in acids but decomposed by fusion with sodium carbonate. Before the blowpipe it is infusible; after moistening with cobalt nitrate solution it turns blue upon intense ignition (as do also kyanite and some other aluminum minerals).

Andalusite is especially characteristic of the contact-metamorphic zones surrounding intrusive igneous rocks, such as granite. It occurs also in regionally metamorphosed tracts. It is produced by the alteration of slates and shales, as described on a later page. Rarely, it occurs in granite as the result of "contamination," i.e., the granite magma dissolved some aluminous sedimentary rock and the excess alumina thus acquired caused andalusite to form along with the normal minerals of the granite.

KYANITE

Kyanite, or cyanite, is generally in long bladed crystals, which rarely show distinct end faces, or in coarsely bladed columnar masses. It is triclinic. It has one perfect *cleavage* (parallel to the front pinacoid a) and another less so (parallel to b); the angle between these is about 74°. The *color* is white to pure blue; the center of a blade may be blue bordered by white margins or rarely, gray or green to black. The *streak* is whitish. It is transparent to translucent with a vitreous to pearly *luster. Hardness* varies in different directions from 5 to 7; it is softer (5) on the front pinacoid in the direction parallel to the length of the crystal; harder (7) in the transverse direction. The *specific gravity* is 3.56 to 3.67. Its *chemical composition* is Al_2SiO_5, and its other chemical and blowpipe properties are like those of andalusite, mentioned above.

Kyanite is a mineral characteristically developed in regions subjected to intense metamorphism. It occurs in gneisses and in mica schists. In the latter the mica is mostly muscovite and in part, possibly, the soda-bearing variety, paragonite. Kyanite is generally associated with garnet and staurolite, or with corundum.

Kyanite is easily distinguished from other minerals, especially andalusite, which has the same chemical composition, by its form, color, variable hardness, specific gravity, and other properties.

Sillimanite. With andalusite and kyanite may be mentioned sillimanite, a mineral of chemical composition identical with theirs, but separated from them because it has a different crystals form as shown by its angles. It can sometimes be recognized with the eye or lens, mostly in gneisses or quartzites, as slender, white or light-colored, four-sided prisms, or radiated aggregates of prisms forming brushes. Its blowpipe and other chemical characters are like those described for andalusite and kyanite.

Andalusite, kyanite, and sillimanite, as already pointed out, have identical chemical composition: they are three polymorphic forms of Al_2SiO_5. Geologically significant is the fact that each was formed under genetically different conditions.

TOURMALINE

Tourmaline is a mineral of which there are many varieties distinguished by color, which in turn depends on the chemical composition. The chief ones are black, green, brown, and red, but the black variety (known also as *schorl)* is the only one important in megascopic petrography.

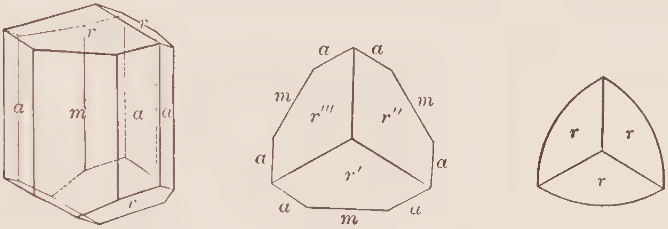

Fig. 35 Fig. 36 Fig. 37

Tourmaline.

Form. Tourmaline crystallizes in the rhombohedral division of the hexagonal system. The faces are therefore in threes or multiples of three. A simple form is shown in Fig. 35; the same form, as it appears to an observer looking down upon the upper end is in Fig. 36. It consists of the three-cornered prism *m*, its edges beveled by the prism faces *a* and terminated by the rhombohedron *r*. The crystals are more complex than this, if well developed, because more faces are present; if both ends are perfect they have unlike faces. Rarely short and thick, the crystals are mostly long thin prisms. Generally the prism faces *a* and *m* alternate repeatedly, so that the prism is striated or channeled, as shown in Fig. 38, and the outline and appearance from above is as shown in Fig. 37. This cross section resembling a spherical triangle is very characteristic of the prisms of rock-making tourmaline. Tourmaline rarely occurs as formless grains or large shapeless masses.

The slender prisms and needles are commonly assembled together into bundles, sheaves, and radiate groups.

General properties. Tourmaline has no *cleavage*, and its fracture is rather conchoidal. It is brittle. The *color* is coal-black, the *luster* glassy, sometimes dull; the *streak* is uncolored and hence not characteristic. The mineral is opaque. The *hardness* is 7 to 7.5; the *specific gravity* 3.1 to 3.2. Tourmaline becomes electrified by friction.

FIG. 38. Tourmaline, showing vertical striae.

Chemical composition. Tourmaline is a complex silicate of boron and aluminum, whose composition is expressed, as a result of X-ray analysis, by the formula $WX_3B_3Al_3(AlSi_2O_9)_3(O, OH, F)_4$, in which $W = Na, Ca$; $X = Al, Fe'', Li, Mg$.

Blowpipe and chemical characters. Tourmaline fuses with difficulty before the blowpipe with swelling and bubbling. If mixed with powdered fluorite and bisulfate of potassium, it momentarily colors the flame a fine green distinctive of the element boron. This valuable test satisfactorily distinguishes tourmaline from the many other minerals for which it may be mistaken. It is decomposed by fusion with sodium carbonate. It is not acted on by acids, but after fusion it gelatinizes in hydrochloric acid.

TABLE 8

ANALYSES OF TOURMALINES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	Na_2O	B_2O_3	H_2O	XyO[1]	Total
I...	35.0	34.4	1.1	12.1	1.8	2.0	9.0	3.7	0.6	99.7
II...	35.6	25.3	0.4	8.2	11.1	1.5	10.1	3.3	4.3	99.8

[1] XyO represents small quantities of other oxides, etc.

I. Paris, Maine.
II. Pierrepont, New York.

Occurrence. Tourmaline is of interest and importance because it is the most common and typical mineral that is produced in the pneumatolytic stage of igneous rock formation, as described on page 144. This mode of origin is indicated by the boron, hydroxyl, and fluorine that it contains. Thus it is abundant in the pegmatite dikes associated with intrusive masses of granites. Its presence in granite indicates, as a rule, nearness to the contact, and it is very likely to occur in rocks that have been altered by contact metamorphism. It is not infrequently associated with certain ore deposits. It appears also in gneisses, in schists, and in crystalline limestones of the metamorphic rocks; its occurrence in these indicates that the metamorphism was induced in part by the consolidating igneous masses giving off water vapor and other volatile substances. The beautiful red and green transparent tourmalines valued as gem material occur in pegmatite dikes of granite, often associated with the common black variety. The red tourmaline is usually associated with lepidolite, the lithium mica.

Determination. The coal-black color, crystalline form, and mode of occurrence of common tourmaline as a rule suffice to identify it. From black hornblende it is easily distinguished by its lack of good cleavage, by its superior hardness, and especially by the shape of the cross section of its prism; and this determination can be made certain by the flame test for boron.

TOPAZ

Topaz crystallizes in the orthorhombic system and the *form* in which it is generally seen is in pointed prisms, as illustrated in Fig. 39. It has a perfect *cleavage* parallel to the base *c*, at right angles to the prism; the *fracture* is uneven. The mineral is very *hard* (8), and brittle. The *specific gravity* is about 3.5. It is generally transparent, colorless, rarely yellow to brown yellow or white and translucent. The *luster* is vitreous. The *chemical composition* is $Al_2SiO_4(F, OH)_2$. Before the blowpipe it is infusible. If topaz is fused in a closed tube with previously fused and powdered phosphorus salt, hydrofluoric acid will be given off and will etch the glass and deposit a ring of silica on the colder upper walls of the tube. If the pulverized mineral is moistened with cobalt nitrate solution and intensely heated before the blowpipe on charcoal, it assumes a fine blue color, thus showing the presence of aluminum.

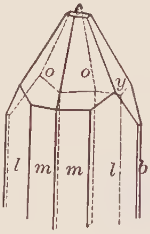

FIG. 39. Topaz.

Although topaz is not a common or important rock-forming mineral, it is highly interesting, as it is particularly characteristic of the pneumatolytic stage in the formation of igneous rocks.

It occurs in crystals in the miarolitic cavities of granites where the vapors have collected during the solidification of the granite, and it occurs in the same way in felsite lavas (especially in rhyolite). It occurs also in pegmatites and in the fissures of the surrounding rocks that served as channel ways for the hot escaping gases, as explained under the descriptions of pegmatites and of contact metamorphism. It is generally associated in these occurrences with quartz, mica, and tourmaline and sometimes with cassiterite and tin ore.

The form, color, cleavage, and great hardness of topaz, together with its mode of occurrence, serve to distinguish it readily from other minerals, and the determination can be confirmed by the chemical tests mentioned above.

Chondrodite

Chondrodite is really one of a small group of minerals so closely allied in all of their general properties that megascopically they are indistinguishable and can all be comprised under this heading. Although chondrodite is monoclinic, it rarely shows as a rock component any definite crystal form that is of value in determining it, but appears as imbedded grains and lumps. The *cleavage* is not marked but may be distinct in one direction. It is brittle with a subconchoidal fracture. The *color* is yellow, honey yellow to reddish yellow, or brown red. The *luster* is vitreous, the *hardness*, 6 to 6.5, and *specific gravity* 3.1 to 3.2. In *chemical composition* chondrodite is closely allied to olivine but differs in containing fluorine and hydroxyl; it is a magnesium fluosilicate—$Mg_5(SiO_4)_2(F, OH)_2$.

As in olivine, the magnesium is generally replaced in part by ferrous iron. The powdered mineral is slowly dissolved by hydrochloric acid, yielding gelatinous silica. The solution, evaporated to dryness and then moistened with acid and taken up in water, yields tests for iron with excess of ammonia after the silica is filtered off, and for magnesium with sodium phosphate solution after separation of the iron by filtering. From olivine it can be distinguished by a test for fluorine as described under topaz. Before the blowpipe it is nearly infusible.

The characteristic mode of occurrence of chondrodite is in dolomite that was subjected to the contact-metamorphic action of igneous rocks. In this it forms yellowish or reddish imbedded grains associated with other contact-metamorphic minerals, such as pyroxene, vesuvianite, magnetite, spinel, and phlogopite. The presence of the hydroxyl and fluorine indicates that chondrodite was formed by pneumatolytic processes.

Appearance, color, mode of occurrence, and associations generally suffice to identify chondrodite, and the identification can be confirmed by the chemical tests mentioned.

B. OXIDES, ETC.

The most important rock-making oxides are silica (SiO_2) and corundum (Al_2O_3). Then come the oxides of iron, which are important as accessory minerals present in nearly all rocks and are therefore widely distributed. Two other minerals not oxides—pyrite (FeS_2) and apatite, the phosphate of calcium—are also included here because of their wide distribution. Goethite ($FeO{\cdot}OH$), which is generally secondary, is placed with the other iron oxides for the sake of convenience.

QUARTZ

Form. Quartz formed below 573° C and at ordinary pressure crystallizes in the trigonal division of the hexagonal system, commonly as a six-sided prism terminated by three positive rhombohedral faces (*r* faces) and three negative rhombohedral faces (*z* faces); the *z* faces are generally much smaller than the *r* faces. This form, known as *low-quartz*, is common as crystals in veins and is illustrated in Fig. 40. In porphyries, in which quartz has formed above 573° and may have crystallized free as phenocrysts, it tends to take the form of a hexagonal bipyramid, as shown in Fig. 41. The prism is rarely developed, but if it is developed it is very short. Quartz formed above 573° is known as *high-quartz*. The *c*-axis in high-quartz is an axis of six-fold symmetry, and hence all six faces are of the same size. In low-quartz the *c*-axis is an axis of three-fold symmetry, and hence the *r* and *z* faces are unlike in size; in fact, some of the *z* faces are small almost to the point of vanishing. When high-quartz cools down below 573°, it transforms instantaneously into low-quartz, but its external crystallographic form is preserved. When low-quartz is heated to 573°, it transforms to high-quartz, but it changes on cooling back to low-quartz. These properties, especially the fact that the crystallographic habit of quartz formed above 573° differs notably from that formed below 573°, make it possible to use quartz as a geologic thermometer. For example, rocks that contain quartz showing trigonal symmetry were formed at temperatures below 573°,

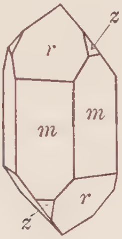

Fig. 40. Crystal of low-quartz.

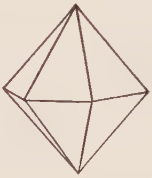

Fig. 41. Crystal of high-quartz.

and those containing quartz of hexagonal symmetry were formed above 573°.

In general, quartz has no definite shape in rocks, especially in igneous rocks such as granite, in which, because it was generally the last mineral to crystallize, its shape was prescribed for it by the adjacent minerals that had already formed. In granites, therefore, quartz is commonly in irregularly shaped grains, molded as it were around the feldspar crystals. In pegmatite it may occur in curious scriptlike figures, as seen on cleavage surfaces of the feldspars with which it is intergrown; this striking intergrowth is known as graphic granite.

Cleavage and fracture. Quartz has so poor a cleavage that for practical purposes it can be regarded as not having any. It has commonly a good conchoidal fracture, which is a great help in distinguishing it in granitic rocks, but in some massive forms the fracture is uneven and splintery. The mineral is brittle to tough.

Color and luster. Rock-making quartz is white through shades of gray and dark smoky gray or brown to black. The gray and smoky tones are most common in igneous rocks, and the white color is most common in sedimentary and metamorphic rocks, but no absolute rule governs this. Black is rare and is mostly confined to igneous rocks. Some quartz has a strong bluish tone. Colorless, limpid quartz, common as crystals in veins and geodes, is rare as a rock-making component but does occur, as in certain very fresh lavas. Quartz may also have an exotic color given it by some substance acting as a pigment; thus it may be red from included ferric oxide dust or green from scales of chlorite, and in the sedimentary and metamorphic rocks, such as quartzite, it may be very dark from included organic matter or charcoal-like substance.

The *luster* varies from glassy to oily or greasy. The *streak* is white or very pale colored and is not distinctive. The *hardness* is 7. Quartz scratches feldspar and glass but is not scratched by the knife. The *specific gravity* is 2.65.

Composition. The crystallized common rock-making quartz is pure silica (SiO_2), but certain massive varieties of silica, such as jasper and opal, which are not crystallized or not apparently so, are of common occurrence, and take part in forming rocks, contain in addition more or less combined water, while impurities like clay and oxides of iron are generally present and give them distinctive colors.

Blowpipe and chemical characters. Quartz is infusible before the blowpipe; varieties that are dark from organic matter whiten but do not fuse. Fused with sodium carbonate, it dissolves with effervescence of carbon dioxide gas. In the sodium metaphosphate bead a fragment floats without dissolving. It is insoluble in acids except hydrofluoric (HF).

Occurrence. Quartz is one of the commonest of all minerals and is universally distributed, occurring in igneous, sedimentary, and metamorphic rocks. Not only does it form rocks in company with other minerals, chiefly feldspar, but in pure sandstones and quartzites it is the only one present. It is indeed so common that, except in the limestones and marbles and dark, heavy igneous rocks such as peridotite, gabbro, and basalt, its presence in rocks should always at least be suspected.

Determination. The hardness of quartz, its lack of cleavage, its conchoidal fracture, and its glassy luster are characters that help to distinguish it, especially from the feldspars with which it is commonly associated. The gray and smoky color it commonly has in granites and other igneous rocks helps in the same way. It can be confused with nepheline, but nepheline is ready soluble in acids with gelatinization. These characters and the blowpipe and chemical properties mentioned above readily confirm the determination of quartz.

Opal and chalcedony. Silica, besides forming the crystallized anhydrous mineral quartz, occurs in noncrystalline amorphous masses containing variable amounts of water. According as the color, structure, and other properties differ, many varieties are produced, which have received particular names. For descriptions of them the larger manuals of mineralogy should be consulted. They appear to have been formed, in large part at least, by deposition from liquids containing colloidal silica, which has been deposited in an amorphous, hydrated condition—as a gel—instead of as crystalline quartz. The resultant mineral is known as *opal*. Its composition is given by the formula $SiO_2 \cdot nH_2O$; its water content ranges from 1 to 20 per cent. By dehydration and crystallization, opal may change to chalcedony. The gelatinous form of silica is illustrated by the product obtained when a silicate such as nepheline is dissolved in an acid and the resulting solution is evaporated.

Opal is also formed in nature as a secretion from water by living organisms, chiefly by diatoms, radiolaria, and siliceous sponges.

Opal is not a primary rock component in igneous or metamorphic rocks, but in the sedimentary rocks it is of considerable importance.

Chalcedony has long been a mysterious mineral but has now been demonstrated to be a fibrous and microfibrous form of quartz, in which the fibers have grown in the direction of the lateral crystallographic axes instead of the vertical axis as in normal quartz. It contains some entangled opal, which accounts for the small but variable amount of water it contains as shown by chemical analyses.

Chalcedony is of waxy appearance. It is generally spherulitic and occurs in botryoidal and mammillary masses; that is, it has typical colloform structures, believed to be inherited from the time it was deposited as gelatinous silica. The *specific gravity* of chalcedony from Yellowstone National Park is 2.55 to 2.61, *i.e.*, lower than quartz, which is 2.65. Chalcedony is common the world over as fillings of amygdules in lavas and is a principal constituent in many cherts.

CORUNDUM

Form. Corundum crystallizes in the hexagonal system and the form assumed is either a thick six-sided prism, swelling out in the middle into a barrel-like shape, or a six-sided tablet. Corundum also occurs commonly in grains or shapeless lumps. The thick and barrel forms are most common when it occurs in massive rocks such as the syenites, and they occur in association with the grains and lumps. On some parting faces a multiple twinning, resembling that illustrated as occurring on feldspars, can be seen; it is produced, however, by another method.

Cohesion properties. Corundum does not have a cleavage, but it has a parting that simulates a perfect cleavage parallel to the base and also in three other directions at an angle to it (parallel to the unit rhombohedron). In large pieces of corundum these partings, or pseudo-cleavages, appear to be nearly at right angles.

Color, luster, and hardness. Rock-making corundum is dark gray, bluish gray, or smoky. It is very rarely blue, forming the variety sapphire, and the red variety, ruby, is excessively rare. The *luster* is adamantine to vitreous but may be dull and greasy in rock grains. Corundum is translucent to opaque. It is the hardest of rock minerals (9). It is brittle, though sometimes very tough. The *specific gravity* is 4.

Blowpipe and chemical characters. Before the blowpipe corundum is infusible. The powder, moistened with cobalt solution and intensely ignited, turns bright blue, showing that aluminum is present. It is insoluble in acids. Its composition is pure alumina (Al_2O_3).

Occurrence. Corundum is an important primary mineral in the igneous rocks of a few regions: in nepheline syenites of Ontario, especially in the pegmatites associated with the syenites, and in other igneous rocks in the Ural Mountains, in South Africa, and in California. An unrecorded occurrence is in syenite in Orange County, New York State. Many more such will doubtless be discovered. In South Africa corundum-bearing pegmatites are the basis of an important mining industry for corundum as an abrasive material. Sapphire, the gem variety of corundum, occurs in basaltic rocks in Yogo Gulch, Montana, the Rhine district, and elsewhere. Corundum occurs also in some contact zones of igneous rocks, where it is usually in thin tabular crystals. It occurs also in metamorphic rocks, in some places as thick beds of the variety called emery.

Determination. The crystal form, if present, and the color are diagnostic properties of corundum, and the identification can readily be confirmed by a test for hardness, since corundum cannot be scratched by any other rock mineral. These tests can be confirmed by the other described properties.

THE IRON OXIDES AND SULPHIDES

Of the iron ores the only ones that can be considered of general importance because of their wide distribution and common occurrence as components of rocks are the oxides and sulfides of iron. Even these play only a minor role in rock-making and are essentially accessory minerals, except where they have become concentrated by geologic processes into considerable masses. They are considered accessory minerals because they occur in small quantities through most rocks of all three great classes of rocks. They include *magnetite, ilmenite, hematite, limonite*, and *pyrite*. There are other oxides and sulfides of iron, but they are of relatively small importance as rock-making constituents.

MAGNETITE

Form. Magnetite crystallizes in the isometric system, most commonly in octahedrons, as in Fig. 42; rarely in dodecahedrons (Fig. 43); and rarely in a combination of both (Fig. 44). It may occur as distinct crystals in rocks but is generally in small grains, the form of which cannot be distinguished. It rarely occurs in larger irregular masses.

General properties. Magnetite has no *cleavage*. Some, however, has a parting parallel to the octahedral faces, resembling cleavage; this is a pressure effect, as it has been experimentally produced. The *fracture* is uneven. Magnetite is brittle with a dark-gray to iron-black, opaque *color*. The *luster* is metallic, fine to dull. It resembles bits of iron or steel in the rocks and is strongly magnetic. The *streak* is black, the *hardness*, 6, and the *specific gravity*, 5.18. The *chemical composition* is $Fe_3O_4 = Fe''O{\cdot}Fe'''_2O_3$; the metallic iron content is 72.4 per cent. It is difficultly fusible before the blowpipe and in the oxidizing flame becomes nonmagnetic. It is slowly soluble in hydrochloric acid.

Occurrence. Magnetite is one of the most widely distributed of all minerals. It occurs in igneous rocks of all kinds, generally in small grains but in some places aggregated into considerable masses. It occurs also in rocks produced by contact metamorphism and in the crystalline schists, sometimes in large bodies. Though ubiquitous in the unmetamorphosed sedimentary rocks, it is not abundant in them. It is not likely to be confused with any other mineral except ilmenite.

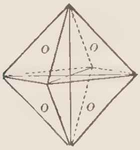

FIG. 42

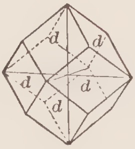

FIG. 43

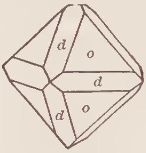

FIG. 44

Magnetite.

Weathering. Magnetite is strongly resistant to weathering. It may alter to limonite, and under the conditions of lateritic weathering it may alter to pseudomorphs consisting of ferric oxide, called martite.

In a few localities immense tabular masses of magnetite cut through the rocks; they are interpreted as dikes formed by the injection of molten magnetite, but this interpretation is far from proved. The great iron-ore body at Kiruna, Sweden—the world's largest body of high-grade iron ore—is regarded as an outstanding example.

Magnetite has formed, under our eyes as it were, at the orifices of some of the fumaroles in the Valley of Ten Thousand Smokes, Alaska. Steam and iron chlorides have reacted at a temperature of 550° C or higher as follows:

$$2FeCl_3 + FeCl_2 + 4H_2O \rightleftharpoons Fe_3O_4 + 8HCl$$

SPINELS

Magnetite can be regarded as the type of a group of minerals known as the spinels. They have the general chemical composition $R''O \cdot R'''_2O_3$ and crystallize in isometric octahedrons, as illustrated in magnetite. In them the RO is MgO, FeO, MnO, ZnO, or mixtures of them; R_2O_3 is Fe_2O_3, Al_2O_3, Cr_2O_3, or mixtures of them. True spinel is $MgAl_2O_4$ ($MgO \cdot Al_2O_3$) and if transparent and of good color is cut as a gem. Hercynite is iron spinel ($FeAl_2O_4$), and chromite is $FeCr_2O_4$, in which more or less other spinel molecules proxy. Depending on their composition, the spinels are black, green, red, or gray. They are extremely hard (7 to 8), without cleavage, and of high luster to pitchy. Some of the spinels are constituents of igneous rocks, especially of those low in silica and rich in iron and magnesia, like peridotite and dunite; others occur in metamorphic rocks, especially in those rocks produced by contact metamorphism. Spinels are only minor components of the rocks and, except in some contact-metamorphic rocks, are rarely found in crystals sufficiently large to make them of megascopic importance.

ILMENITE

General properties. Ilmenite crystallizes in the hexagonal system, but it is so rarely in good megascopic crystals in rocks that its crystal form is not important. It is usually as embedded grains and masses, or as plates of irregular to hexagonal outline. It is brittle and has no *cleavage*. *Fracture* is conchoidal. The *color* is iron black, rarely with a faint reddish to brownish tinge; the luster is submetallic; the *streak*, black to brownish red. Ilmenite is opaque with a *hardness* of 5.5 to 6 and a *specific gravity* of 4.7. The *composition* is $FeTiO_3 = FeO{\cdot}TiO_2$. It is generally not pure but is more or less mixed with hematite (Fe_2O_3), with which it is isomorphous. Before the blowpipe it is very difficultly fusible; in the reducing flame it becomes magnetic. After fusion with sodium carbonate it can be dissolved in hydrochloric acid, and the solution, after addition of hydrogen peroxide, becomes yellow, thus showing titanium. Fresh mineral is difficultly soluble in acids; it is decomposed by fusion with potassium bisulfate. The solutions give reaction for iron with potassium ferricyanide. The test for titanium is the safest method to determine the mineral.

Occurrence. Ilmenite, or titanic iron ore, as it is often called, is widely distributed as an accessory component in igneous rocks in the same manner as magnetite, which it generally accompanies. It occurs also in gneisses and schists as a widespread minor component. Unless the imbedded grains are of such size that they can be safely tested, ilmenite cannot usually be distinguished from magnetite by simple inspection. The most important megascopic occurrences are in the coarser-grained gabbros and anorthosites, where ilmenite is very common and, in fact, occurs in places in such large masses, notably in the Adirondacks and Quebec, that it is a commercial source of titanium.

HEMATITE

Form. Hematite crystallizes in the rhombohedral division of the hexagonal system, but it is so rarely in distinct, well-formed crystals of observable size as a rock constituent that crystal form is not a matter of practical importance.

It occurs as a rock-mineral in three different guises: *specular iron ore, micaceous hematite,* and *common red hematite.*

As specular iron ore it forms masses and plates, the latter sometimes hexagonal in outline. Its *color* is black to steel gray, rarely with a faint reddish tone. It is opaque, and has a metallic *luster,* which as a rule is brilliant or splendent, so that it resembles polished steel; at other times it is rather dull but metallic looking. The *fracture* is subconchoidal. There is no *cleavage.*

As micaceous hematite it is in thin flakes that somewhat resemble mica; they may be so thin as to be translucent and they are then deep red. The *luster* is submetallic to metallic but may be splendent like that of the specular form. The thin leaves generally have ragged outlines, but some are hexagonal.

Common red hematite does not seem to be crystallized, but X-ray study has proved it to be minutely crystalline. It is massive, columnar, granular, or in stalactitic or mammillary masses, and some is earthy (red ocher). It is dull, without metallic *luster,* opaque, and dark red.

General properties. The *streak* of hematite is red, from bright Indian red to brownish red, and furnishes the most convenient method of distinguishing hematite from magnetite and limonite. Before the blowpipe hematite is very difficultly fusible except in fine splinters. After being heated in the reducing flame, it is magnetic. It dissolves slowly in hydrochloric acid, and the solution reacts for iron.

The *composition* is Fe_2O_3, ferric oxide; metallic iron content is 70 per cent. The *hardness* ranges from 5.5 to 6.5; but the red ocherous variety, because of its fine state of division, seems to be much softer than 5.5. The *specific gravity* of the specular variety is 5.26.

Occurrence. Hematite is one of the most widely distributed of minerals. Micaceous hematite, often called specularite, is a common accessory component of feldspathic igneous rocks, such as granite. It occurs abundantly in megascopic form in certain crystalline schists, as in itabirite, which is a quartz-hematite schist. Also, in minute microscopic scales it forms the red coloring matter in igneous and metamorphic rocks. The red color of many potassium feldspars is due to it and so is that of many slates.

Common red hematite occurs as beds and masses in sedimentary and metamorphic rocks, many of them of great size, and constitutes one of the most valuable ores of iron. It is the interstitial cement of many stratified rocks, such as red sandstones, and as a red pigment in the form of powder it is everywhere distributed in rocks of all classes and in soils, though some of the pigment may be turgite (hydrohematite), which closely resembles it.

Limonite and Goethite

Definition. Limonite until recently was thought to be an amorphous form of ferric hydrate. X-ray investigation, however, has shown that there is little or no amorphous limonite—it is chiefly microcrystalline goethite ($HFeO_2$). The formula of goethite corresponds to a water content of 10.1 per cent, but the water content of many limonites is much higher and has a wide range. This additional water is interpreted as held by capillary entanglement in the microfibrous goethite and by admixed, amorphous silica. "Limonite" still remains a convenient field term.

Form. Limonite occurs in earthy masses in the rocks, and when in considerable deposits frequently shows stalactitic, mammillary shapes, and concretionary forms.

General properties. Limonite has no *cleavage*. The *luster* of compact varieties is silky to submetallic, but generally dull and earthy. The *color* is a shade of brown, from very dark to brownish yellow. The surface of the compact stalactitic or mammillary forms may have a varnish-like skin. It is opaque. The *streak* is yellow brown. The *hardness* of the compact mineral ranges from 5 to 5.5, and the *specific gravity* from 3.6 to 4.0. Difficultly fusible before the blowpipe, it becomes magnetic in the reducing flame. Heated in closed glass tube, it gives off much water. It is slowly soluble in hydrochloric acid, the solution giving reactions for iron. The yellow streak is the most convenient means of distinction from hematite.

Occurrence. Limonite occurs in several different ways. It is commonly secondary, formed at the expense of previously existing minerals as the result of weathering. It is frequently seen as small, earthy, yellowish to brownish masses that represent the oxidized remnants of some previous iron-bearing mineral, such as pyrite, siderite, glauconite, or hornblende. As bog iron ore it is loose, porous, and earthy. Mixed with more or less clay it forms yellow ocher. Limonite is the yellow pigment of many soils and sedimentary rocks.

Pyrite

Form. Pyrite occurs almost invariably as crystals in the rocks, rarely as grains and masses. It crystallizes in the isometric system. It is frequently seen in cubes or in the twelve-sided form shown in Fig. 45, called the pyritohedron because pyrite so commonly shows it.

Combinations of the cube and pyritohedron are also common, as shown in Fig. 46. The cubic faces are generally striated by fine lines, as seen in Fig. 47, produced by oscillatory combination of the pyritohedron on the cube faces. The octahedron is far less frequent; it is likely to be modified by the pyritohedron, as shown in Fig. 48. Other more complex forms also occur.

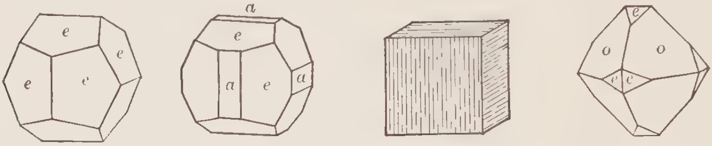

FIG. 45 FIG. 46 FIG. 47 FIG. 48

Pyrite.

General properties. There is no *cleavage*. The *fracture* is conchoidal to uneven, the *color*, brass yellow, and the *luster*, metallic and splendent, duller when tarnished. Pyrite is opaque. The *streak* is brownish black. The *hardness* is notably high for a sulfide: 6 to 6.5. The *specific gravity* is 5.0 and the *composition*, FeS_2 (iron = 46.6 per cent; sulfur = 53.4 per cent. It is easily fusible before the blowpipe, burning and giving off sulfur dioxide gas, and leaving a magnetic globule. When heated in the closed glass tube it gives a sublimate of sulfur and leaves a magnetic residue. It is insoluble in hydrochloric but decomposes in boiling nitric acid with separation of sulfur.

The color and crystallization usually suffice at once to identify pyrite and to distinguish it from other minerals. From pyrrhotite ($Fe_{11}S_{12}$) and chalcopyrite ($FeCuS_2$), other sulfides of iron occasionally seen in rocks, it is readily distinguished by the test for hardness: chalcopyrite (3.5) can be readily scratched with the knife and furthermore gives reactions for copper; pyrrhotite also is scratched by the knife and in addition has a bronze color, is magnetic, and gives ittle or no sulfur in the closed tube.

Occurrence. Pyrite has many different modes of origin and consequently occurs in rocks of all kinds as a scattered component, generally in small distinct crystals, less commonly aggregated. The largest masses are found in ore deposits, chiefly formed by the action of hydrothermal solutions. In igneous rocks it occurs in minute amounts as a primary product of crystallization from the molten magma. In sedimentary rocks it is common as replacements of fossils, and its occurrence in this way must be due to reactions between the sulfur of decaying albuminous matter of organic origin and the iron in the inclosing sediments. It is common in coal beds, forming the so-called "coal brasses."

APATITE

Apatite crystallizes in hexagonal prisms either rounded at the ends or capped by a six-sided pyramid. It is scratched by the knife, has a vitreous *luster*, and is white, green, or brown. There is no good *cleavage*. It is brittle, transparent in small crystals to opaque in large, and very difficultly fusible. It dissolves in nitric acid, and ammonium molybdate solution added to a few drops of the nitric acid solution gives a bright yellow precipitate, showing the presence of phosphorus. The *composition* is phosphate of calcium with fluorine: $(CaF)Ca_4(PO_4)_3$; chlorine may proxy *wholly* or in part for the fluorine.

Apatite is found in large, even huge, crystals in pegmatites and in metamorphosed limestones in the crystalline schists; these may be said to be its chief megascopic modes of occurrence. In these, however, apatite cannot be said to function as a rock-mineral of wide or general importance. In addition, however, it occurs as minute microscopic crystals, which can seldom be detected with the eye or lens, in all kinds of igneous rocks and in many metamorphic ones. Microscopical study of the thin sections of such rocks has shown that in this form apatite is nearly universally distributed as an accessory component.

Although the proportion of apatite is small, rarely exceeding two or three-tenths of one per cent of the rock in which it occurs, its presence is a matter of great importance, literally of vital importance, since it supplies to the soil that is formed when the rocks decay by weathering the phosphorus so necessary to vegetable and animal life (in bones and teeth).

Section 2. HYDROUS SILICATES

The hydrous silicates are of secondary origin: they are formed from previously existent minerals by the agents of weathering, water containing carbon dioxide or vegetable acids, and heated water or its vapors circulating in already solid, existent rocks. Therefore the hydrous silicates play no important part in fresh, unchanged igneous rocks; only as these rocks alter do the hydrous silicates become abundant in them. Their true home is in the metamorphic and sedimentary rocks, which at times are made up wholly of these minerals.

The important minerals considered in this section are *kaolinite*, the *chlorites*, *serpentine*, *talc*, and the *zeolites*. Some micas would also naturally be considered here, and, among secondary minerals, limonite also, but, for reasons previously stated, these have been treated in the foregoing section.

KAOLINITE

General properties. Kaolinite crystallizes in the monoclinic system, forming small thin plates or scales, generally with hexagonal outlines; these plates are flexible and recall mica but are inelastic. The scales are generally so minute and so closely aggregated together, however, that the crystal form is no help in megascopic determination. Kaolinite commonly occurs in compact, friable, or mealy masses. The *color* is white but may become yellow, brown, or gray. Neither the hardness (2 to 2.5), nor the *specific gravity* (2.6) can be used for practical tests. Rubbed between the fingers kaolinite has a smooth, unctuous, greasy feel, which helps to distinguish it from fine aggregates of some other minerals occurring in nature.

It is infusible before the blowpipe, but if moistened with cobalt nitrate and ignited, it turns blue, showing the presence of aluminum. Heated in the closed glass tube, it yields water. It is insoluble in hydrochloric acid. In the phosphorus bead before the blowpipe, undissolved silica is left; this test helps to distinguish it from *bauxite,* a mixture of hydrous oxides of aluminum, which much resembles it and in places occurs in large deposits. Bauxite dissolves in the phosphorus bead completely. Chemically, kaolinite is a hydrous silicate of aluminum, $Al_2Si_2O_5(OH)_4$. This is often written $H_4Al_2Si_2O_9$, or $Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$ ($Al_2O_3 = 39.5$ per cent; $SiO_2 = 46.5$ per cent; $H_2O = 14$ per cent).

Occurrence. Kaolinite is a secondary mineral formed by the weathering of previously existent aluminous silicates, chiefly feldspar. The reaction by which it is formed from the alkali feldspar is one of the most important in nature, for by it soil is chiefly made, and the alkali necessary for plant food is liberated and converted into soluble form. It probably takes place as follows:

$$\overset{\text{Orthoclase}}{2KAlSi_3O_8} + \overset{\text{Water}}{2H_2O} + \overset{\text{Carbon dioxide}}{CO_2} \rightleftharpoons \overset{\text{Kaolinite}}{H_4Al_2Si_2O_9} + \overset{\text{Potassium carbonate}}{K_2CO_3} + \overset{\text{Quartz}}{4SiO_2}$$

This process and reaction have already been described under feldspars. Under other conditions, as already mentioned, the feldspars may alter to other so-called clay minerals or to gibbsite.

Every stage in the change of feldspar to kaolinite can be seen in nature, as described in the chapter dealing with the origin of sedimentary rocks and soils. Thus kaolinite is intimately mixed with the feldspar substance of such rocks as are undergoing this change. It occurs in places as extensive deposits where such rocks were completely altered in place and the products of the alteration remain where they were formed; it also occurs in extensive beds in the sedimentary formations. As the flakes of kaolinite are very minute, light, and flat, they remain much longer in suspension than the other products of land waste, and thus in erosive and sedimentary processes they continuously tend to become separated from the particles of other minerals. We find beds of kaolinite showing every degree of admixture with sand, etc., and grading into sandstones and other rocks, but some beds of kaolinite are of a high degree of purity.

CHLORITES

The chlorites are a numerous group of micaceous silicates so named because of their green color (Greek *chloros*, green). They are mostly secondary in origin, formed at the expense of previously existing silicate minerals that contain aluminum, iron, and magnesium. Outwardly they resemble the micas, but differ from them in that their folia are not elastic. They have certain properties in common by which they can be recognized as a group; but to identify the different members is difficult and for megascopic purposes of little importance. In the description that follows, then, it is the group properties that are given, though these are based largely on the species clinochlore, perhaps the most common member of the group.

Form. The chlorites are really monoclinic in crystallization, but, as in the case of the micas, when they show crystal form they are in six-sided plates and tablets. Generally, however, they occur in irregular leaves and scales, which are massed together in aggregates ranging from finely granular to coarsely massive, less commonly in fanlike or rosette-like groups. The scales may be flat, or in bent and curled forms.

General properties. Chlorite, like mica, has a perfect *cleavage* in one direction parallel to the flat base of the plates. The cleavage leaves are flexible and tough, but, unlike those of mica, they are not elastic. The *luster* of cleavage faces is rather pearly. The *color* is a variable green, usually a rather dark green. Chlorite is usually translucent. Its *hardness* is 2 to 2.5 soft, just scratched by the fingernail. The *specific gravity* is about 2.7, and the *streak* is pale green to white. In *chemical composition* the chlorites are hydrous iron-magnesium-aluminum silicates: they can be illustrated by the formula given to clinochlore, $(Mg, Fe'')_5Al_2Si_3O_{10}(OH)_8$.

Before the blowpipe chlorites are not fusible or are very difficultly so; with the borax bead they react for iron. Heated in the closed glass tube, they yield water. They are insoluble or soluble with difficulty in hydrochloric acid but are decomposed in sulfuric acid. These reactions are those given by the common kinds of chlorite.

Occurrence. The chlorites are abundant and occur wherever previously existent rocks containing magnesian silicates, such as biotite, amphibole, and pyroxene, have been altered by geologic processes. To their content of chlorite many igneous rocks owe their green color, the original ferromagnesian silicates having altered more or less completely into this mineral.

Chlorite is abundant also in many schistose rocks. In chlorite schist it is the prominent megascopic component; and other finer-grained foliates owe their green color to its presence, as, for example, green slates. Thus as finely disseminated particles it is a common coloring matter.

Some of the iron-rich chlorites occur abundantly in the sedimentary iron-ore beds.

SERPENTINE

General properties. Serpentine occurs in an amazing variety of forms and colors. It is usually massive, ranging from minutely to coarsely granular. Some is fibrous, the fibers being fine, flexible, and easily separable. Massive varieties have a conchoidal or splintery fracture and a smooth, greasy feel. The *color* of massive varieties is green, bright yellowish green, olive green, blackish green, or nearly black. The fibrous varieties are generally brownish, yellowish brown, pale brown, or nearly white. The *luster* of the massive varieties is greasy, waxlike, glimmering, and usually feeble to dull; the luster of fibrous varieties is pearly to opalescent. Serpentine is translucent to opaque. The *hardness* is 2.5 to 3.0; an apparent greater hardness may be caused by presence of remains of the original mineral or by infiltrated silica. The *specific gravity* is somewhat variable: for the fibrous varieties, 2.2 to 2.4; for the massive, 2.5 to 2.7. The *composition* is $Mg_3Si_2O_5(OH)_4$ ($MgO = 43.0$ per cent, $SiO_2 = 44.1$ per cent, $H_2O = 12.9$ per cent). A small part of the magnesium is generally proxied by ferrous iron. Before the blowpipe, serpentine is difficultly fusible; fine fibers fuse more readily. In the closed tube, it yields water on ignition. The finely powdered material decomposes in boiling hydrochloric acid with separation of silica but does not gelatinize. The solution may be tested for iron and magnesium.

Serpentine is easily distinguished from epidote and other common green silicates that may resemble it, by its greasy feel and softness.

Occurrence. Serpentine is a secondary mineral formed by the alteration of previously existing silicates rich in magnesium. Thus pyroxene, amphibole, and especially olivine are commonly found to be altered to serpentine. The process with olivine can be illustrated by the following equation:

$$\overset{\text{Olivine}}{2Mg_2SiO_4} + \overset{\text{Water}}{2H_2O} + \overset{\text{Carbon dioxide}}{CO_2} = \overset{\text{Serpentine}}{Mg_3Si_2O_5(OH)_4} + \overset{\text{Magnesite}}{MgCO_3}$$

The equation suggests why magnesite ($MgCO_3$) is commonly associated with serpentine. If magnesite is not present, it may have been dissolved by the carbonated water and removed from the locus of alteration.

Another method of alteration to serpentine is by the action of hydrothermal solutions containing dissolved silica:

$$3Mg_2SiO_4 + H_2O + SiO_2 = 2Mg_3Si_2C_5(OH)_4$$

Therefore, as an alteration product of such minerals, especially as the result of the action of hydrothermal solutions, serpentine is a common and widely distributed mineral and occurs both in igneous and in metamorphic rocks. It is disseminated in small masses in the rocks or forms large independent bodies, as described further under the chapters dealing with the rocks.

Besides the common massive form, many varieties of serpentine are known. The most important of these is the finely fibrous variety, the asbestos of commerce, which is known as *chrysotile.* It occurs as veinlets in rocks (serpentinites) composed chiefly of massive serpentine. Bright-green massive material is known as precious serpentine and is cut for ornamental purposes.

TALC

General properties. Talc rarely occurs in distinct crystals. It is usually in compact or strongly foliated masses, or in scaly or platy aggregates grouped into globular or rosette-like forms. Like mica, it has a perfect *cleavage* in one direction, but the resulting laminae, though flexible, are not elastic; it is sectile. It has a smooth, greasy feel. The cleavage face has a mother-of-pearl *luster.* The *color* is white, often inclining to green, apple green, or, sometimes, gray to dark gray. It is usually translucent. The *hardness* is 1 to 1.5, easily scratched with the fingernail. The *specific gravity* is 2.7 to 2.8. The *streak* is white. Talc marks a cloth, the marking being best seen on dark cloth: this feature is the single distinctive test for talc. The *composition* is hydrous magnesium silicate: $Mg_3Si_4O_{10}(OH)_2$. Before the blowpipe talc whitens, exfoliates, and fuses with difficulty on the edges. It yields water in the closed glass tube only on intense ignition; it is scarcely acted on by hydrochloric acid. It is easily recognized by the properties mentioned above. When in fine scales, however, it cannot be distinguished from sericite except by chemical tests.

Occurrence. Talc is a secondary mineral produced by the action of hot circulating fluids on magnesium silicates, especially those free from aluminum, such as olivine, hypersthene, and some pyroxenes and amphiboles. The process is illustrated by the following equation.

$$\overset{\text{Enstatite}}{4MgSiO_3} + \overset{\text{Water}}{H_2O} + \overset{\text{Carbon dioxide}}{CO_2} = \overset{\text{Talc}}{Mg_3Si_4O_{10}(OH)_2} + \overset{\text{Magnesite}}{MgCO_3}$$

Thus talc occurs in the igneous rocks as an alteration production of such silicates, especially in the peridotites and pyroxenites. The place, however, where it is particularly abundant is in the metamorphic rocks, where it forms large masses, as in steatite or soapstone, or it may be an important component of several varieties of schistose rocks, as in talcose schists.

ZEOLITES

The zeolites are a group of hydrous silicates, composed, like the feldspars, of aluminum with alkali and alkali-earth metals. They are mostly secondary minerals formed at the expense of feldspars and feldspathoids by the action of hot circulating waters and steam, and consequently they are found chiefly in igneous rocks, especially in volcanic rocks. They form a group not so closely related in crystallization and other properties as the feldspars, but still they have many properties in common by which they can be distinguished. The properties will first be described, and then the individual characters of a few of the most abundant zeolites will be presented.

Group properties. The zeolites are generally well crystallized, the crystals presenting the forms characteristic of the different species. They have a vitreous luster; most are colorless or white, but some are yellow or red. Most are of inferior hardness and can be scratched by the knife. Their specific gravity is low, 2.1 to 2.4. They fuse very readily before the blowpipe, most of them with marked intumescence (whence the name, from the Greek, *zein*, to boil); some, however, fuse quietly, to white glasses or enamels. They dissolve in hydrochloric acid, some of them gelatinizing and some yielding slimy silica. The more common zeolites are *analcite, natrolite, stilbite,* and *heulandite.*

Analcite. This zeolite crystallizes in isometric trapezohedrons like those of garnet, a property that enables one easily to recognize analcite if it is in crystals. It is colorless or white. Before the blowpipe it first becomes opaque, then fuses quietly to a clear glass, coloring the flame yellow. It dissolves in hydrochloric acid with separation of silica but does not gelatinize. Its composition is $NaAl(SiO_3)_2 \cdot H_2O$.

Natrolite crystallizes in orthorhombic prisms, generally long, slender, and even needle-like and arranged in divergent bunches or compacted into fibrous, radiating masses. Before the blowpipe it fuses easily and quietly to a clear glass; it also fuses in a candle flame. It dissolves in acid with gelatinization. The composition is $Na_2Al(AlO)(SiO_3)_3 \cdot 2H_2O$.

Stilbite crystallizes in complex monoclinic crystals, which are generally so aggregated as to form a sheaf. It has a perfect cleavage in one direction, along which stilbite has characteristically a pearly luster. It may occur in divergent, or in globular groups. It is white or, more rarely, red. Before the blowpipe it swells, intumesces, and fuses to a white enamel. It dissolves in acid without gelatinization. The composition is essentially a hydrous calcium-sodium-aluminum silicate: $(Ca, Na)_3Al_5(Al, Si)Si_{14}O_{40} \cdot 15H_2O$.

Heulandite crystallizes in flat monoclinic crystals which aggregate into compound individuals, the crystals having grown side by side with the flat surfaces together. It has a perfect cleavage parallel to this flat side, which has a pearly luster. The cleavage plates are generally curved and have a lozenge-shaped outline. Blowpipe and chemical characters are like those of stilbite. The composition of heulandite is essentially a hydrous calcium-aluminum silicate: $Ca_6Al_{10}(Al, Si)Si_{29}O_{80} \cdot 25H_2O$.

Occurrence. As stated above, the zeolites are secondary minerals occurring chiefly in igneous rocks. They occur where igneous rocks were subjected to the action of steam or hot circulating waters, which attacked the feldspars and feldspathoids. Hence, feldspathic rocks that are somewhat altered are likely to contain zeolites in small amount scattered through them; in some rare instances the rocks are largely composed of them, especially of analcite. Ordinarily, the zeolites cannot be detected megascopically, but they may be discovered by heating some of the powdered rock in a closed glass tube, whereupon their presence is indicated by the easy evolution of abundant water.

Their especial home is in lavas, particularly basalts. Here they coat the walls of cavities and the sides of joint planes, and they fill the blow-holes in the form of amygdules, as described under amygdaloidal structure and under basalt. They may be associated with crystals of quartz and of calcite in such occurrences. Besides the zeolites already mentioned, many other zeolites may occur. Description of the less common zeolites can be found in the larger manuals on minerals.

Section 3. CARBONATES

The important rock-making carbonates are calcite and dolomite.

Calcite

Form. Calcite crystallizes in the rhombohedral system. Its crystals as a rule are well developed and perfect, and some are of large size. It has a great variety of crystal forms, many of them being complex. Some simple crystals are shown in the annexed figures. Fig. 49 shows a simple flat rhombohedron, three faces above and three below. Fig. 50 shows the unit rhombohedron, so called because its faces are parallel to the cleavage. Fig. 51 is of a very acute rhombohedron. Fig. 52 shows a short prism having six prism faces *m* and the flat rhombohedron *e* above and below; Fig. 53 is similar to Fig. 52, but the prism faces *m* are elongated. Fig. 54 is an acutely pointed form, the scalenohedron. All these are crystal forms commonly shown by calcite; they occur where it lines the cavities in rocks, in druses, in amygdaloids, in geodes, and on the surfaces of joint planes and fissures and in caves; in short, in all those places where calcite has been deposited by infiltrating water containing it in solution.

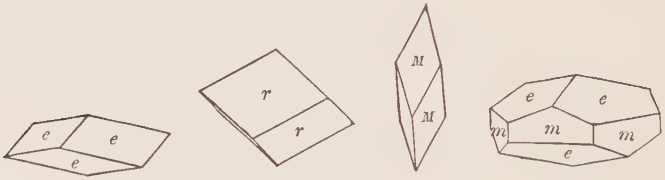

Fig. 49 Fig. 50 Fig. 51 Fig. 52

Calcite.

As a rock-making mineral calcite is massive; it is coarse to finely granular in marble, compact in ordinary limestone, or loose and powdery in chalk. It has an open-work or spongelike structure in travertine and a rounded, stalactitic structure in the dripstone of cave deposits and in concretions; it is not uncommonly fibrous.

Cleavage. Calcite has a perfect rhombohedral cleavage, in three directions parallel to the rhombohedral faces *r* of the crystal shown in Fig. 50. Although this cleavage is of course best produced in isolated crystals, it can be readily seen on the fractured surfaces of coarsely crystalline massive rocks, as in many marbles and related rocks and in the massive calcite of veins. The angles of the face of the rhomb produced by cleavage are 78° and 102°, as shown in Fig. 55, and small cleavage pieces can be readily tested by applying them to the edges of the figure on the paper.

General properties. The *hardness* is 3; calcite is therefore readily scratched or cut by the knife. The *specific gravity* is 2.710, if pure. Calcite is generally colorless or white, but it displays a great variety of exotic colors owing to the presence of impurities; thus it may be reddish or yellowish from iron oxides, gray to black from organic matter, or green, purple, or blue from other substances. The *streak* is white to gray. Crystallized calcite has a vitreous *luster*; massive forms are glimmering to dull. It varies from transparent through translucent to opaque. The *chemical composition* is $CaCO_3$, or $CaO{\cdot}CO_2$ ($CaO = 56$ per cent, $CO_2 = 44$ per cent).

FIG. 53. Six-sided prism of calcite.

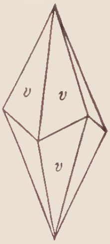

FIG. 54. Scalenohedral calcite.

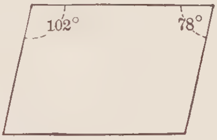

FIG. 55. Face of cleavage rhomb of calcite.

Before the blowpipe it is infusible, colors the flame reddish yellow, and if placed on moistened yellow turmeric paper after ignition colors it brown. Fragments effervesce freely in *cold* (differing thus from dolomite) very dilute acids.

Occurrence. Calcite is one of the most abundant and widely distributed of all minerals. It occurs in the igneous rocks as the result of alteration of the calcium-bearing silicates by waters containing dissolved carbon dioxide, other substances being formed as byproducts at the same time. The calcite thus formed may remain for a time in the rock, but eventually, as the latter breaks down into soil, it is to a greater or lesser extent carried away in solution.

Calcite occurs also in minute cavities in unaltered igneous rocks, especially intrusive ones, and its origin is probably due to infiltration and deposition of the material from neighboring rock masses. In these cavities the calcite is ordinarily not observable megascopically, but its presence is easily ascertained by immersing a fragment of the rock in cold dilute acid and seeing if it effervesces and gives off carbonic acid gas. Calcite occurs also in amygdaloidal cavities in lavas, especially in basalts, generally in good crystals.

In the sedimentary and metamorphic rocks calcite is much more important. It is very commonly distributed through them either in fine particles or forming a cement to the other mineral granules. From this minor role, it increases more and more in abundance as a constituent, until finally there are enormous, widely extended rock masses, such as the chalks, limestones, and marbles, composed largely or wholly of calcite. Such rocks are described in their appropriate places in this work; it is sufficient here to mention that in the sedimentary rocks calcite is an important constituent of chalk, limestone, calcareous marls, and calcareous sandstones; in travertine, dripstones, and veins; and in the metamorphic rocks in marbles and in rocks that are mixtures of calcite and various silicates.

Determination. Calcite, if sufficiently coarsely crystallized, is easily recognized by its inferior hardness and rhombohedral cleavage. This determination is confirmed chemically by its ready solubility in cold dilute acids with effervescence of carbon dioxide gas, and if necessary by a test for the presence of calcium. For the distinction from dolomite, reference should be made to the description of that mineral.

Dolomite

Form. Dolomite crystallizes in the rhombohedral system and, like calcite, occurs as simple rhombohedral crystals the faces of which are parallel to the cleavage, as in Fig. 50 of calcite. Unlike that mineral, it rarely forms complex crystals, and the simple rhombohedron, which is the prevalent habit when dolomite shows crystal form, usually has curved faces, as represented in Fig. 56, instead of flat. Moreover, the curved crystals are likely to be compound, that is, made up of several subindividuals. This is the habit dolomite has where it lines druses and cavities. As a rock-making mineral, however, dolomite is generally massive and ranges from coarsely granular to exceedingly fine grained.

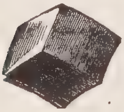

Fig. 56. Dolomite crystal.

Cleavage. The cleavage, like that of calcite, is perfect in three directions parallel to the faces of the simple rhombohedron. The angles of the cleavage rhombs (74° and 106°, nearly) differ only a degree or so from those of calcite, and accordingly they cannot be distinguished from those of calcite by the unaided eye.

General properties. The natural *color* is white, and though this is common, dolomite is likely to be tinted some exotic color by other substances; thus it may be reddish, brown, greenish, gray, or even black. The *luster* is vitreous, pearly to dull or glimmering in compact varieties. Dolomite is translucent to opaque. The *hardness* is 3.5 to 4.0, harder than calcite but easily scratched with a knife. The *specific gravity* of the pure mineral is 2.87. The *chemical composition* is $CaMg(CO_3)_2$ ($CaO = 30.4$ per cent, $MgO = 21.9$ per cent, $CO_2 = 47.7$ per cent). Before the blowpipe it is infusible, but when placed after ignition on moist turmeric paper it colors it brown.

If unpowdered, dolomite reacts feebly in cold dilute acid and does not dissolve, but if powdered it effervesces briskly even with cold acid Even lump dolomite effervesces in hot acid and dissolves rapidly. The solution can be tested for calcium and magnesium as directed in the chapter on mineral tests. The difference in action of cold acid on powdered dolomite as contrasted with its effect on the unpowdered is a convenient means of distinguishing dolomite from calcite in the field.

Occurrence. Dolomite is a scattered accessory component of certain crystalline schists and in beds of gypsum. Its great importance as a rock-making mineral lies in the fact that it forms thick, areally extensive beds both in the sedimentary series and in the metamorphic rocks. That is, there are rocks that are composed wholly of dolomite; these rocks, like their constituent mineral, are called dolomites. Thus dolomite exactly parallels calcite, and in limestones and dolomites (and in the marbles derived from them) we have every degree of association between the two minerals—rocks that are composed of calcite alone, those that have both in varying percentages, and those that are pure dolomite. These are described in more detail under the carbonate rocks.

Determination. The rhombohedral cleavage and inferior hardness separate dolomite, like calcite, from other common rock minerals. The frequent curved surfaces help to distinguish it from calcite, but the test with acid already mentioned, together with the finding of magnesium in the solution, is the only safe way.

Siderite, magnesite, and breunnerite. As an appendix to the description of calcite and dolomite, these carbonates, which are of local importance, may be mentioned. Siderite is ferrous carbonate ($FeCO_3$), magnesite is magnesium carbonate ($MgCO_3$), and breunnerite is an isomorphous mixture of the two $(Mg, Fe)CO_3$. In crystallization, cleavage, hardness, and other properties, they closely resemble the carbonates described. Siderite is usually light to dark brown, magnesite is white, and breunnerite is brownish. Siderite occurs chiefly more or less massive and impure in certain sedimentary deposits, as the so-called "clay ironstone," and is a valuable ore of iron. Magnesite occurs chiefly in association with serpentine and talc. It may be accompanied or replaced by breunnerite.

Section 4. SULFATES

As rock-making minerals, only two sulfates of the many known are abundant: *gypsum* and *anhydrite*. Barite ($BaSO_4$) is common in veins and occurs also in concretions, but it does not form rock masses or play a role as a rock component as gypsum and anhydrite do.

Gypsum—Selenite

Form. Gypsum crystallizes in the monoclinic system, and the common form of the crystal is shown in Fig. 57. The same crystal is shown in Fig. 58 revolved so that the side pinacoid face *b* is parallel with the plane of the paper; such crystals can be roughly tested by placing them on the diagram and seeing if the angles coincide. Twin crystals are common; they tend to assume arrow-head forms, as shown in Fig. 59. More commonly, as a rock constituent, gypsum is granular; less commonly it is foliated, with curved surfaces; some is fibrous.

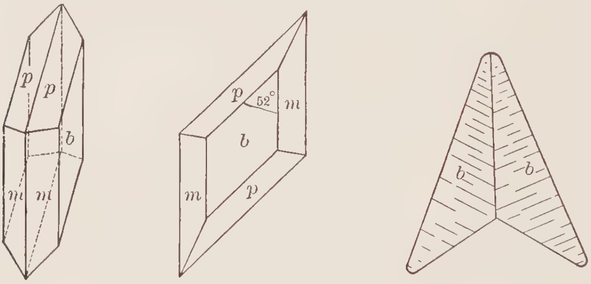

Fig. 57 Fig. 58 Fig. 59

Some forms of gypsum crystals.

Cleavage. Gypsum has a perfect cleavage parallel to the side face *b*; by means of it very thin sheets with perfect luster can be split off from good material, almost as in mica. Such sheets break in one direction in straight lines with a conchoidal fracture; this is due to another cleavage parallel to the edge between the prismatic faces *mm*. If such sheets are bent, cracks will appear in them, making angles of 66° to 114° with the straight fracture edge mentioned above; if the bending parallel to this direction is continued, the plate will break with a fibrous fracture, and a cleavage rhomb like that shown in Fig. 60 will be obtained. In massive coarsely crystalline gypsum, these cleavages can usually be obtained, and they furnish one means of helping to identify it; in compact massive forms it is likely that no cleavage can be seen. Fibrous gypsum simply cleaves parallel to the fibers.

General properties. Gypsum is generally colorless or white, and the crystals are transparent to translucent, but some is reddish or yellowish; massive varieties may even be red, brown, or black, because of the presence of impurities, and translucent to opaque. The *luster* of the cleavage face is glassy to pearly; that of fibrous varieties is satiny, whereas massive forms are glistening and glimmering to dull. The *streak* is white. The *hardness* is 1.5 to 2.0, easily scratched by the fingernail. The *specific gravity* of pure crystals is 2.32. The *chemical composition* is hydrous sufate of calcium: $CaSO_4{\cdot}2H_2O$ ($CaO = 32.6$ per cent, $SO_3 = 46.5$ per cent, $H_2O = 20.9$ per cent). Before the blowpipe it fuses easily, and after ignition it colors moistened turmeric paper brown. Fused with sodium carbonate and charcoal dust on charcoal and transferred to a moistened surface of silver, it stains it dark. Finely powdered, it dissolves readily in boiling dilute hydrochloric acid. Heated intensely in a closed glass tube, it gives off water and becomes opaque. Heated moderately (not above 200°) it loses some water and becomes plaster of Paris, and the powder, if moistened, takes up water again and sets (becomes solid), turning back into gypsum. If heated too highly, gypsum loses all its water and becomes anhydrite ($CaSO_4$); this is called dead-burnt plaster because it does not set.

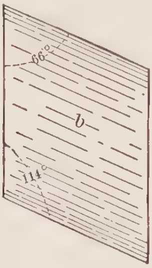

FIG. 60. Gypsum cleavage rhomb.

The occurrence of gypsum is mentioned later in its description as a stratified rock.

ANHYDRITE

General properties. Anhydrite crystallizes in the orthorhombic system. In the rocks in which it occurs it is granular, ranging from coarse to extremely fine grained; less commonly it is fibrous. It has *cleavages* in three directions at right angles, and in coarsely crystalline anhydrite this cleavage can be seen to produce cubelike forms. Most is white, but some is tinted as in gypsum. The *luster* of the cleavage faces is pearly to glassy; in massive varieties it varies to dull. Anhydrite is harder (3 to 3.5) than gypsum but is easily cut by knife.

The *specific gravity* is 2.95. The *chemical composition* is $CaSO_4$ ($CaO = 41.2$ per cent, $SO_3 = 58.8$ per cent). Blowpipe and other reactions are as with gypsum, but it does not yield water in the closed glass tube on heating, which is the best distinction from gypsum; the pseudo-cubic cleavage also differs from that of gypsum.

Occurrence. Like gypsum, anhydrite forms beds that are interstratified in sedimentary formations, especially in limestones and shales. It is also found in masses and in geodes in such rocks and is a common associate of rock salt and gypsum.

ROCK SALT—HALITE

Rock salt is sodium chloride ($NaCl$); it is the chief chloride that occurs as a rock-forming constituent in such amounts as to be important. It is easily recognized by its cubic crystals, perfect cubic cleavage, solubility in water, and saline taste. Colorless and transparent to white, and translucent; it is generally tinted various colors by impurities. The hardness is 2.5.

It occurs in beds, some of which are of enormous extent, in the sedimentary formations, and is generally accompanied by gypsum and anhydrite, and in places by other valuable chlorides, such as sylvite (KCl) and carnallite ($KMgCl_3 \cdot 6H_2O$). More remarkable even than the great salt beds are the intrusive plugs and stocks of salt up to 40 square miles in area that occur in the Gulf Coast regior of Louisiana and Texas, in Germany, Iran, and other places.

CHAPTER V

DETERMINATION OF ROCK MINERALS

The more important physical properties of the rock minerals have been described in the preceding chapter, and the methods by which most of the minerals can be determined have been stated. The present chapter gives qualitative chemical tests, which can generally be made with a few reagents and simple apparatus, and which are of great help in determining minerals and consequently in determining rocks. These tests are followed by two tables: one for rough approximations in the field, the other for complete identifications in the laboratory by means of the properties and methods described.

CHEMICAL TESTS

Chemical tests consist mainly in observing the effect of acids on the mineral: whether it is dissolved, only partly attacked, or wholly insoluble; if it is soluble, or partly so, the solution is treated with certain reagents in order to ascertain what substances have gone into solution. These reagents serve to detect the acid radicals and the metals that compose the minerals. A few useful additional tests are given.

A. Powdering the sample. The first thing to do in testing the solubility of minerals is to prepare a finely powdered sample. Small chips, grains, or splinters of the mineral, about the size of wheat grains and as pure as possible, are successively crushed and ground to a fine powder, like flour, in a steel or agate mortar, until a sufficient amount has been prepared. The fragments are best crushed in a steel mortar, and unless iron is to be tested for they can be ground in the mortar as well. The finer the powder is ground, the more readily does it dissolve; therefore it is generally best to grind the powder until it no longer feels gritty when a small pinch is rubbed between the fingers.

B. Treatment with acid. A small bulk of the powder prepared as directed, about equal to a pea in volume, is put into a test tube and covered with an inch of distilled water, and a few drops of nitric or hydrochloric acid are added. Either acid may be used, but if the solution is to be tested subsequently for phosphorus or chlorine, nitric acid should be employed. If cold acid has no apparent effect on the substance, the test tube should be gently heated over the flame of a Bunsen burner or lamp of a suitable kind until the liquid is brought to the boil. If the effect is slight or inappreciable, more acid is added, and the boiling is repeated until the mineral dissolves or it becomes apparent that it cannot be dissolved.

C. Carbonic acid—carbonates. If the mineral effervesces briskly in cold dilute acid, it is a carbonate and carbon dioxide gas is being given off. The common rock-making carbonates are calcite (carbonate of calcium) and dolomite (carbonate of calcium and magnesium). Calcite effervesces freely when cold acid acts on an uncrushed fragment and more violently on a powder. On the other hand, dolomite, unless reduced to a powder, will not effervesce with cold acid; however, when a fragment is treated with hot acid, it effervesces briskly, and this effervescence will of course be greatly accelerated if the dolomite is pulverized. This difference of behavior is a convenient means of distinguishing between calcite and dolomite; the identification can be confirmed by tests for calcium and magnesium as given below. The rare rock mineral cancrinite, a silicate containing carbon dioxide, also effervesces very slowly in hot dilute acid; the hot solution should be examined in a good light with a lens, when a slow persistent evolution of minute bubbles will be seen. In observing the hot solution care must be taken not to confuse the ebullition of steam bubbles with the effervescence of carbon dioxide gas; a moment's pause should be given to allow the former to cease.

D. Soluble silicates. Gelatinization. If the mineral treated according to A dissolves wholly or partly without effervescence, it should be tested for soluble silica. If it is only partly attacked, the insoluble portion should be filtered off and the filtrate used. This is boiled down in the test tube, which should be continuously and gently shaken to prevent cracking, until the solution is greatly concentrated, if necessary to a few drops. It is then allowed to cool and stand, and if it becomes a jelly the presence of soluble silica is indicated. If the amount of silicate that has dissolved is relatively large, a jelly will probably form while the solution is being boiled down and is still hot; otherwise the solution must be concentrated and allowed to cool, as just stated. Care must be taken not to confuse the true jelly of soluble silica with a thickening of the cooled solution owing to concentration in it of the salts, especially basic salts of iron. If the solution is carefully evaporated to dryness and the residue is heated for a few moments, the salts, on being moistened with strong hydrochloric acid and then warmed, will dissolve in water, whereas the silica will be left as an insoluble residue and can be filtered off. The filtrate can be tested for the metallic bases—aluminum, iron, calcium, etc.—which may be in solution, by the methods described below.

The rock-making silicates that dissolve on boiling with nitric or hydrochloric acid are nepheline, sodalite, analcite, olivine, chondrodite, serpentine, anorthite, leucite, noselite, stilbite, heulandite, and cancrinite. All except serpentine, leucite, analcite, stilbite, and heulandite yield gelatinous silica; and when the liquid in which these have been dissolved is boiled it turns from the milkiness caused by suspended material to a translucent appearance as the result of the slimy silica suspended in it. The bases, however, have gone into solution.

E. Insoluble silicates. Fusion. Most rock-making silicates are insoluble in acids or only partly soluble. To get them into solution so that they can be tested for the bases as described in the following sections, they must first be fused with sodium carbonate (Na_2CO_3). For this purpose some of the powder obtained in A is mixed with 4 to 5 times its weight of dry anhydrous sodium carbonate, placed in a platinum crucible or spoon, and gently heated to redness over a Bunsen burner. If no crucible is at hand, a coil of platinum wire can be used instead; the mixed powder is made into a thick paste with a little water, and a quantity is taken on the coil and carefully fused before the blowpipe. A fused bead the size of a large pea can thus be obtained. The fusion must be continued until bubbling ceases.

In this fusion the silicates are decomposed, and silica is liberated and takes the place of the carbonic acid in the sodium carbonate, which is thus converted into sodium silicate. The liberated carbon dioxide is given off with bubbling and frothing of the fusion, and this effect in itself indicates the presence of silica in the original substance, provided that previous trial has shown that the frothing does not come from combined water.

The fused mass obtained is broken up in a mortar, placed in a test tube, and treated with acid until dissolved; it is then evaporated, the silica is separated, and the metallic bases are brought into solution just as directed for soluble silicates in D.

If the fusion was made in a platinum crucible, the resulting cake can generally be loosened and removed by boiling with a little water; if not, it is dissolved with water and acid in the crucible, which is set in a beaker or dish.

F. Aluminum. The filtrate from the silica obtained in D or ED combined can be tested for aluminum. It is heated to boiling after addition of a few drops of nitric acid, and ammonia is added in slight excess. If a white or light-colored, flocculent, gelatinous precipitate forms, it is aluminum hydroxide. If it is reddish brown, it is wholly or in part ferric hydroxide, indicating the presence of iron, which is likely to be present in silicates, especially colored ones, and the aluminum if present may be masked by it.

If much magnesium is present in the mineral, its hydroxide is likely to be precipitated at this point by the ammonia, unless the solution is rather dilute and much ammonium chloride or nitrate has been formed by the neutralization of the acid by the ammonia.

Some of the precipitate is scraped from the filter paper and transferred to a clean test tube; if it is small in amount, it is transferred paper and all; it is then covered with about 5 cubic centimeters of water; a piece of pure caustic potash (KOH) about the size of a pea is dropped in, and the whole is boiled. The aluminum, if present, will dissolve, leaving the iron hydroxide undissolved; the latter can now be filtered off. The filtrate should be made slightly acid with hydrochloric acid and boiled, and ammonia should be added in slight excess; aluminum, if present, will be precipitated as the white flocculent hydroxide.

Aluminum can also be detected before the blowpipe by intensely heating on charcoal the powdered mineral, moistened with cobalt nitrate. The presence of aluminum is indicated by the mass turning blue, as mentioned under topaz, kyanite, etc., in the following tables.

G. Iron. The detection of iron has been mentioned above in F. A more delicate method is to add a few drops of potassium ferrocyanide solution to a few drops in water of the final filtrate obtained in D or ED combined, after boiling with a few drops of nitric acid if hydrochloric was originally used. The formation of a deep-Prussian-blue precipitate (or coloration if the solution is very dilute) indicates the presence of iron. The nitric acid converts the ferrous iron in the solution into the ferric state. Potassium *ferro*cyanide produces a Prussian blue with *ferric* salts, not with ferrous, while conversely potassium *ferri*cyanide produces the same effect with *ferrous* salts. Thus by testing portions of the original solution of the mineral in hydrochloric acid with these two reagents, the state or states of oxidation of the iron in the original mineral can be ascertained.

Iron is also shown to be present if minerals become magnetic after being heated in the reducing flame of the blowpipe.

H. Calcium. The ammoniacal filtrate from the hydroxides of aluminum and iron obtained in F, or the clear liquid if ammonia failed to cause precipitation, may contain calcium salts. To prove the presence of calcium the liquid should be heated to boiling and some ammonium oxalate added, whereupon the formation of a precipitate of calcium oxalate proves the presence of calcium. If the solution is further to be tested for magnesium, the calcium oxalate must be removed by filtration; it is allowed to stand for some time and then filtered; if the filtrate runs through turbid it should again be passed through the paper until the liquid is clear. To this a little more ammonium oxalate is added to prove the complete precipitation of the calcium.

I. Magnesium. Ordinarily the test for magnesium should not be applied until the aluminum, iron, and calcium have been removed, as directed in F and H, or their absence has been ascertained. To the solution thus obtained some sodium phosphate and a considerable quantity of strong ammonia are added. The formation of a precipitate, ammonium magnesium phosphate, proves the presence of magnesium. If a precipitate does not form at once, it is not safe, however, to consider magnesium absent, for if the amount is small and the solution is warm, the precipitate may not appear until the liquid has become cold and has stood for some time. It is then likely to appear as a crystalline precipitate attached to the walls of the vessel.

J. Sodium. A mineral containing sodium, when heated before the blowpipe, colors the flame bright yellow. The best effect is obtained with silicates if the powdered mineral is mixed with an equal volume of powdered gypsum and a little of this mixture is taken on a clean, moistened platinum wire which has been previously tested for cleanliness. The reaction is so delicate, however, and is produced so strongly by minute quantities of sodium or accidental traces of it that very good judgment must be used in observing it. Only if the coloration is intense and prolonged can the presence of sodium as an essential constituent be inferred.

K. Potassium. Potassium can be detected by the violet color it imparts to the Bunsen or blowpipe flame. From a silicate the flame color is best obtained by powdering the mineral and mixing it with gypsum, as mentioned under sodium in J. The flame color is delicate and is entirely obscured by any sodium present; it can however be seen if it is viewed through a piece of thick, dark-blue glass which cuts off all but the potassium flame. Through this glass the potassium flame will appear violet or reddish purple.

Another test is to take a small portion of the final filtrate obtained in D or ED combined, evaporate it to a very small volume, add an equal volume of alcohol, and, if it is turbid, filter it. Then a few drops of hydrochlor-platinic acid (H_2PtCl_6) are added, and if a heavy yellow or orange-colored crystalline precipitate, potassium platinic chloride (K_2PtCl_6) forms it shows the presence of potassium. No ammonium salt must be present, for it will yield a similar precipitate.

L. Hydrogen—water. If a little of the powdered mineral is placed in a glass tube, one end of which has been closed by fusion and drawing off, and gently heated below redness, the evolution of water, which collects on the upper walls of the tube, shows that it contains loosely attached *water of crystallization.* This phenomenon occurs with zeolites, such as analcite ($NaAlSi_2O_6 \cdot H_2O$), and with gypsum ($CaSO_4 \cdot 2H_2O$). On the other hand, some minerals, including many silicates, contain hydrogen and oxygen firmly attached in the form of *hydroxyl* (OH), which is given off as water only at a very high heat. Indeed some minerals, as for instance staurolite and talc, $Mg_3Si_4O_{10}(OH)_2$, must be heated white hot before the blowpipe before the water is given off. This difference in behavior often serves as a useful test in determining minerals. Many minerals containing hydroxyl also contain fluorine, and the water evolved in the tube from such fluorine-bearing minerals gives an acid reaction to test paper and the glass is likely to be etched. Unless the glass is etched, however, the test is not decisive evidence of the absence or presence of fluorine.

M. Fluorine. The best test for fluorine is, as described under topaz, on page 63, in a bulb tube with sodium metaphosphate.

N. Chlorine. Chlorine occurs in rock salt, apatite, and sodalite. The test for chlorine is its precipitation as silver chloride (AgCl) by addition of a few drops of silver nitrate. The white precipitate turns bluish gray on exposure to light. The test for chlorine is very delicate, and slight impurities may cause a faint opalescence in the liquid on addition of the silver salt. Rock salt is easily told by its solubility in water, taste, and associations. Apatite generally contains only a little chlorine, yielding therefore a faint test, or it may contain no chlorine. Sodalite dissolves in dilute nitric acid, and on adding silver nitrate a considerable precipitate of the chloride is obtained. The nitric acid solution yields also gelatinous silica as in D; these tests suffice to identify sodalite.

O. Sulfuric acid. Barium chloride produces in a solution containing a sulfate a heavy white precipitate of barium sulfate ($BaSO_4$), insoluble in hydrochloric or nitric acid. Gypsum, anhydrite, and noselite contain sulfuric acid; they dissolve in hydrochloric acid, and the solution may be tested as above. Noselite yields also gelatinous silica, as in D, and the two reactions serve to identify it.

P. Phosphoric acid. The powdered mineral (see A) is dissolved in nitric acid, and a little of a solution of ammonium molybdate is added, whereupon a yellow precipitate of ammonium phosphomolybdate shows the presence of phosphorus. This test is very delicate. It should be made in a cold or nearly cold solution. The precipitate is soluble in excess of ammonia. If it is desired to make this test in a mixture of minerals, as in a fine-grained rock for instance, and if silica may be in the solution, it is best to evaporate the solution and get rid of the silica as directed in D. The test for phosphoric acid can then be applied to the filtrate acidified with nitric acid. Apatite is the only common rock-making mineral that contains phosphoric acid, and its presence in rocks and soils can generally be shown by this test when it cannot be detected megascopically.

TABLES FOR THE MEGASCOPIC DETERMINATION OF ROCK MINERALS

The two tables that follow will be found useful in identifying the commoner rock-making minerals. Besides the minerals given in these tables, there are many that are less common in rocks but of some local importance, for instance, the minerals in the metamorphic limestones and schists.

The more important minerals have been described in the preceding chapter on the characters of minerals, but only about fifty minerals or mineral groups comprising the kinds ordinarily met in megascopic rock study are included here. The tables can be used only to distinguish from one another the minerals that are named in them; they cannot be used to distinguish these minerals from all other minerals. If doubt arises and a mineral seems not to be one of those described here, larger manuals of descriptive and determinative mineralogy should be consulted for the identification.

Table 1. This table is based entirely on obvious and easily determinable physical properties. It includes about thirty common minerals or mineral groups. It can often be used to advantage in the field. The only equipment required is a lens, a pocket knife, and fragments of quartz and feldspar, in addition to the hammers usually carried. It is of advantage to have one blade of the knife magnetized. The streak or color of the powdered mineral can be tested by grinding a small piece to powder between two hammer faces, pouring it on a piece of white paper, and rubbing the dust with the finger to observe the color produced. A mineral fragment can be cracked into smaller grains, and these grains can be examined with a lens to observe the cleavage if it is not well shown by the mineral on the fractured rock surface. The transparency or translucency, if not obvious in the mineral inclosed in the rock, can be tested by holding a fragment or sliver against the light and observing whether light is transmitted through its thinnest edges. The hardness is best tested on a smooth,

lustrous cleavage face with the knife point or a sharp-pointed fragment of quartz or feldspar, substances as a rule readily obtainable.

Table 2. This table includes about fifty prominent rock minerals or mineral groups, the characters of which are given in the foregoing descriptive portion. It requires for its use some of the simpler apparatus and reagents available in every chemical and mineralogical laboratory, and the knowledge that enables one to use them. The apparatus and reagents have already been mentioned on page 8.

The table is based on tables of the Brush-Penfield Determinative Mineralogy, which have been modified to meet the demands of this particular book, and if further information is desired that manual can be consulted to advantage.

Table 2 is much more nearly complete and certain in its identification than Table 1 and should be used in preference when possible. Table 1 is a more or less rough method of approximation to be used in the field or when no apparatus or reagents are at hand.

The tables, it must be repeated, cannot be used in identifying all minerals that occur in rocks but only to distinguish those that are mentioned in them. The identification of a mineral is generally complete, but it may happen that some comparatively rare mineral will give similar reactions. Thus the mineral aragonite falls into the same place in the table as calcite, but reference to the description of calcite shows at once that it differs markedly in other properties, such as cleavage and crystallization. This will usually be found to be so, and if further information is desired it must be sought elsewhere. But within the limits imposed, the tables should serve a useful purpose to the student of rocks.

TABLE 1

The mineral has a fine cleavage in one direction; is micaceous and can be split into thin leaves by the use of the knife point. Sec. 1 below.

Has a good cleavage in two directions. Sec. 2.

Has a good cleavage in three directions, forming cubes or rhombs. Sec. 3.

Has a fine fibrous structure and cleavage. Sec. 4.

No apparent good cleavage. Sec. 5.

Section 1. Cleavage in one direction

A. Micaceous. Cleavage leaves flexible, *elastic*, tough (i.e., without brittleness). Occurs in crystals, shreds, flakes. Black, brown, gray, or white. Transparent-translucent. *Mica*, p. 39.

B. Micaceous. Cleavage leaves tough, flexible, *non-elastic*. In crystals, shreds, masses. Generally green to dark green. *Chlorite*, p. 77.

C. Micaceous as a rule. Leaves flexible but *non-elastic*. Greasy feel, very soft. Marks cloth. White, greenish, gray. Generally in foliated masses. Translucent. *Talc*, p. 79.

D. Leaves somewhat flexible but showing cross cleavage cracks when bent; in one direction fibrous, in the other brittle, forming rhombs. Soft; scratched by finger nail, but not greasy in feel. Generally colorless, white or reddish; transparent to translucent. In crystals, masses, seams. *Gypsum,* p. 86.

E. Leaves have a brilliant metallic luster, like polished steel. *Hematite* (micaceous variety), p. 71.

F. Leaves brittle, lozenge-shaped outline. Generally white, translucent. Scratched by the knife. Crystals in cavities. *Heulandite,* p. 81.

G. Not micaceous, massive, brittle. Very hard, not scratched by knife or feldspar. Yellowish green to dark green. In crystals or masses. *Epidote,* p. 56.

Section 2. Cleavage in two directions

A. Two cleavages at or very nearly at 90°. Brittle; hard, scratched by quartz but not by knife. Generally of a light color, white, pink to red or gray; translucent. In crystals, grains, masses. *Feldspar,* p. 27.

B. Generally of a dark color, greenish to black; in grains or short prisms; may be light colored in metamorphic rocks and there generally elongated columnar in the cleavage direction. Cleavage good but not eminent; prismatic. Cleavage angles 87° and 93°. As a rule can be scratched by feldspar. *Pyroxene,* p. 45.

C. Generally of a dark color, greenish to black. Commonly in crystals elongated or bladed in cleavage direction. Light colored in many metamorphic rocks. Cleavage very good with shining surface. Cleavage angles 55° and 125°. As a rule scratched by feldspar. *Amphibole,* p. 47.

Section 3. Cleavage in three directions

A. Cleavages not at right angles, forming rhombs. Easily scratched by knife. Usually white, sometimes tinted various shades to black; transparent to translucent. In crystals, masses, veins, etc. *Calcite,* p. 82, or *Dolomite,* p. 84. (If rhombic surfaces of crystals are curved, probably dolomite.)

B. Cleavages at right angles forming cubes; soluble, strong saline taste. Transparent; colorless or white, rarely tinted. In crystalline masses. *Halite,* rock salt, p. 88.

C. Cleavage apparently cubic. No perceptible taste. Easily scratched by the knife. White, bluish. In crystalline masses. *Anhydrite,* p. 87.

D. Apparent cleavages forming rhombs, some is seemingly cubic. Very hard, scratches quartz easily. In hexagonal crystals, grains, or lumps of a dark, smoky, or bluish gray; more or less translucent. *Corundum,* p. 68.

Section 4. Fibrous or Columnar structure

A. In opaque brown to black masses. Streak yellow brown. *Limonite,* p. 73.

B. In opaque red-brown to black masses. Streak brownish red. *Hematite,* p. 71.

C. White or reddish; translucent. Brittle. Radially fibrous, or showing slender prismatic crystals. Difficultly scratched by the knife. Occurring in cavities, veins, or seams. *Natrolite,* p. 81.

D. White or reddish; translucent or transparent. Brittle. Radially fibrous groupings common. Compound crystals may be sheaf-shaped. Scratched by knife. In cavities, veins, or seams. *Stilbite,* p. 81.

E. Shreds easily into fine, flexible fibers like cotton or silk. *a.* White or light gray. *Amphibole asbestos,* p. 49. *b.* White to yellowish brown; silky; generally in veins in or associated with serpentine. *Chrysotile* (serpentine) *asbestos,* p. 78.

F. White or pale colors. Translucent. Brittle, easily scratched by knife but not by fingernail. In masses. *Calcite,* p. 82.

G. White to pale red. Silky luster; translucent. Brittle; soft, scratched by fingernail. In masses and seams. *Gypsum,* p. 86.

Section 5. Without good or apparent cleavage

A. Opaque, brass-yellow crystals with metallic luster. Not scratched by the knife. *Pyrite,* p. 73.

B. Opaque, earthy, brown to brown-black masses. Streak yellow brown. Scratched by the knife. *Limonite,* p. 73.

C. Opaque, reddish brown to black masses, or crystals and grains, iron black with metallic luster. Streak brownish red. Scratched by the knife. *Hematite,* p. 72.

D. Opaque, iron-black masses, grains, or octahedrons with metallic luster. Streak black. Magnetic. Not scratched by the knife. *Magnetite,* p. 69.

E. Opaque, black grains or masses with reddish tone. Luster metallic to submetallic. Streak black to reddish black. Not noticeably magnetic. Scarcely or not scratched by the knife. *Ilmenite,* p. 71.

F. In garnet-shaped crystals or spherical. Usually dark red to black and translucent. Brittle. Not scratched by feldspar. *Garnet,* p. 54.

G. In garnet-shaped crystals. Colorless or white to gray white, translucent. Scratched by feldspar but not by knife. *Leucite,* p. 38, or *Analcite,* p. 81.

H. In transparent to translucent crystals or grains of a light yellowish-green or bottle-green color. Not scratched by feldspar. *Olivine,* p. 52.

I. In prismatic crystals, generally slender, shiny, and black with triangular cross section. Not scratched by quartz. *Tourmaline,* p. 61.

J. In grains, masses, or hexagonal, pyramidal crystals. Conchoidal fracture. Greasy to glassy luster. Colorless, white, smoky, dark; transparent to translucent. Not scratched by feldspar. *Quartz,* p. 65.

K. In grains or masses, rarely in crystals having rectangular or hexagonal sections. Conchoidal fracture. Greasy, oily luster. White, gray, or reddish; translucent. Scratched by feldspar. *Nepheline,* p. 37.

L. In grains or masses, generally bright blue. *Sodalite,* p. 37.

M. In masses. Dark or yellowish green. Easily scratched or cut by knife. *Serpentine,* p. 78.

N. In masses, generally somewhat foliated. Greasy feel; very soft, marks cloth. White, greenish, gray. *Talc,* p. 79.

O. In hexagonal crystals, grains, or lumps. Dark smoky or bluish gray; translucent. Very hard, not scratched by quartz, garnet, or tourmaline. *Corundum,* p. 68.

P. In masses, compact or chalky. Friable, very soft, easily cut by fingernail. Rubbed between the fingers, has a soft soapy feel. *Kaolinite,* p. 76.

TABLE 2

1. *A.* Has a metallic luster or is opaque; dark or strongly colored streak. 2.*
   B. Without metallic luster or transparent or translucent on very thin edges; streak white or light colored. 6.

* The appended number refers to that in front of a succeeding section.

2\. *A*. Burns when heated in the blowpipe flame, and gives off sulfurous fumes. Brass-yellow color. *Pyrite*, p. 73.
B. Becomes magnetic when cold, after being heated in the reducing blowpipe flame. Not brassy in appearance. Infusible or very difficultly so. Iron oxides. 3.

3\. *A*. Magnetic without heating. *Magnetite* (and in part *Ilmenite*), p. 69.
B. Magnetic only after heating. 4.

4\. *A*. Gives water when heated in the closed glass tube. *Limonite*, p. 73.
B. Gives little or no water. 5.

5\. *A*. Reacts for titanium. *Ilmenite*, p. 71.
B. No reaction for titanium. Streak brownish or Indian red. *Hematite*, p. 71.

6\. *A*. Fusible before the blowpipe (fusibility 1–5). 7.
B. Infusible or fusible only with difficulty. 17.

7\. *A*. Becomes magnetic after heating before the blowpipe in reducing flame. 8.
B. Does not become magnetic. 11.

8\. *A*. Soluble in hydrochloric acid with separation of silica, sometimes gelatinous. 9.
B. Insoluble in hydrochloric acid or only slightly acted on. 10.

9\. *A*. Micaceous or foliated. *Mica* (Biotite or Lepidomelane), p. 39.
B. Isometric crystals. Gelatinizes imperfectly. *Garnet* (Andradite), p. 55.
C. Gelatinizes. *Olivine*, p. 52. Rich in iron. *Fayalite*, p. 52.

10\. *A*. Micaceous; fusible with difficulty. *Biotite*, p. 39.
B. Isometric crystals or spherical in shape. After fusion gelatinizes with HCl. Dark red color. *Garnet* (Almandite), p. 55.
C. Quietly and difficultly fusible. Greenish black or bronze brown. Good cleavage. *Pyroxene* (Hypersthene), p. 44.
D. Fuses with intumescence, coloring flame strong yellow. Black. Prismatic cleavage angle 55°. *Amphibole* (Arfvedsonite), p. 49.
E. Fuses quietly, coloring flame yellow. Black, prismatic cleavage, 87°. *Pyroxene* (Aegirine), p. 45.

11\. *A*. Readily and completely soluble in water; has a saline taste. *Halite* (rock salt), p. 88.
B. Difficultly soluble in water. After intense ignition, colors moistened turmeric paper brown.
  a. Gives much water in closed glass tube, *Gypsum*, p. 86.
  b. Gives little or no water in closed glass tube, *Anhydrite*, p. 87.
C. Soluble in hydrochloric acid without gelatinizing or separation of silica on evaporation. A drop of sulfuric acid in the solution precipitates calcium sulfate. *Apatite*, p. 75.
D. Soluble in hydrochloric acid with gelatinization.
  a. When heated in closed glass tube, gives off water. 12.
  b. When heated as above, yields little or no water. 13.
Soluble in hydrochloric acid; silica separates, but no jelly forms.
  a. When heated in closed glass tube, gives off water. 14.
  b. When heated as above, yields little or no water. 15.
E. Insoluble in hydrochloric acid. 16.

12. *A.* Fuses quietly to a clear transparent glass. White or colorless; in slender crystals or fibrous bunches. *Natrolite*, see Zeolites, p. 81.
B. A fragment in warm dilute hydrochloric acid gives off minute bubbles of CO_2 gas. *Cancrinite;* see Feldspathoids, p. 36.

13. *A.* Fuses rather easily before the blowpipe, coloring the flame a strong yellow. Dissolves easily in very dilute nitric acid and gelatinizes.
  a. The nitric acid solution gives a precipitate with silver nitrate solution (p. 94). Color usually blue. *Sodalite*, p. 37.
  b. The nitric acid solution gives a precipitate with barium chloride solution. *Hauynite-Noselite*, p. 38.
  c. No reaction with silver nitrate or barium chloride. *Nepheline*, p. 37.
B. Difficultly soluble in hydrochloric acid and colors the flame very little. Has a good cleavage in two directions. *Anorthite*, see Feldspars, p. 25.

14. *A.* Usually in greenish masses, compact, greasy, sometimes fibrous. Fusible with difficulty. *Serpentine*, p. 78.
B. Fuses quietly to a clear glass, coloring flame yellow. Generally in colorless or white garnet-like crystals. *Analcite*, p. 81.
C. Fuses with swelling and intumescence. Commonly in sheaflike or radiated groups of crystals. *Stilbite*, p. 81.
D. Fuses as in *C*. Crystals have a fine cleavage with pearly luster and lozenge-shaped section. *Heulandite*, p. 81.

15. *A.* Fuses quietly to a glassy globule. Slowly acted on by hydrochloric acid. Good cleavage in two directions; one cleavage surface generally shows fine parallel lines (twinning striae). Grayish or bluish; may show a play of colors. *Labradorite;* see Feldspars, p. 25.
B. Fuses quietly to white globule. Easily soluble in hydrochloric acid; when solution is evaporated to dryness and the residue is moistened with a little hydrochloric acid and dissolved in water and filtered, ammonia produces little or no precipitate. Generally white, *Wollastonite* ($CaSiO_3$).

16\. *A*. Micaceous. Cleaves into thin flexible elastic plates in one direction. *Micas*, p. 39.

B. Micaceous. Cleaves into thin plates, flexible but not elastic. Very soft and has a greasy fee. Color white, gray, or greenish. *Talc*, p. 79.

C. Cleavable, micaceous, but cleavage plates not elastic, though flexible. Soft but not so soft as talc. Color green, usually rather dark green. *Chlorite*, p. 77.

D. Not micaceous. Solid and brittle. Good cleavage in two directions at or about 90°. Generally light colored, red, or gray. Hard, cannot be scratched by knife. Fusible with difficulty. *Feldspar*, p. 25.

E. Fuses with swelling and bubbling before the blowpipe. Very hard, scratches feldspar. Generally in coal-black columnar crystals; less commonly brown or green. No cleavage. *Tourmaline*, p. 61.

F. Fuses quietly. Gelatinizes with hydrochloric acid after fusion. Crystals as on p. 54 or in spherical forms. Very hard. No good cleavage. *Garnet*, p. 54.

G. Fuses with swelling and intumescence to a black slaggy mass, which gelatinizes in hydrochloric acid. Powdered mineral on intense heating in closed glass tube yields a little water. Yellowish to blackish green. *Epidote*, p. 56.

H. Fuses quietly or with little intumescence. Generally scratched by feldspar.
  a. Prismatic cleavage with angle of 87°. *Pyroxene*, p. 45.
  b. Prismatic cleavage with angle of 55°. *Amphibole*, p. 47.

I. Fuses with intumescence to a greenish or brownish glass, which will gelatinize with hydrochloric acid. *Vesuvianite* (Idocrase), p. 57.

17\. *A*. After intense ignition before the blowpipe, gives a brown stain when placed on moistened turmeric paper. 18.

B. Dissolves in hydrochloric acid but without gelatinizing or yielding a residue of silica on evaporation. 19.

C. *a*. Dissolves in hydrochloric acid and gelatinizes. *Olivine*, p. 52.
  b. Reacts for fluorine (see Topaz, 22*F*). *Chondrodite*, p. 64.

D. Dissolves in hydrochloric acid; does not gelatinize but silica separates. 20.

E. Insoluble in hydrochloric acid.
  a. Can be scratched by glass or a knife point. 21.
  b. Cannot be scratched by glass or the knife. 22.

18\. *A*. Effervesces freely in *cold* dilute acid. *Calcite*, p. 82.

B. If powdered, effervesces freely in *cold* acid. *Dolomite*, p. 84.

19\. *A*. Becomes magnetic when heated in the reducing blowpipe flame.
  a. Little or no water in closed tube; streak brown red. *Hematite*, p. 71.
  b. Water in closed glass tube; streak yellow brown. *Limonite*, p. 73.

B. In hexagonal crystals chiefly. Gives reactions for phosphorus. A little dilute sulfuric acid gives a precipitate of white $CaSO_4$ in the cold concentrated solution of mineral in hydrochloric acid. Readily scratched by the knife. *Apatite*, p. 75.

20\. *A*. Commonly in compact green masses. Also fibrous, like asbestos; then white or brownish or yellowish. Greasy feeling; easily scratched by knife. *Serpentine*, p. 78.

B. In spherical or garnet-shaped crystals. White to gray. *Leucite*, p. 38.

21.
- *A.* Micaceous. Cleavage leaves tough and elastic. *Micas,* p. 39.
- *B.* Micaceous. Cleavage leaves tough and flexible, but not elastic. Intense ignition in closed tube gives water. Color green. *Chlorite,* p. 77.
- *C.* Very soft and has a greasy feel. *Talc,* p. 79.
- *D.* Claylike, compact or mealy. Leaves undissolved silica in the phosphorus salt bead. Gives water in the closed glass tube. *Kaolinite,* p. 76.

22.
- *A.* Extremely hard. Scratches quartz. Generally has a parting that looks like cleavage. *Corundum,* p. 68.
- *B.* No cleavage; conchoidal fracture. Scratches feldspar. Some as hexagonal crystals with pyramid at end. *Quartz,* p. 65.
- *C.* Prismatic cleavage. Becomes black before the blowpipe, and very fine splinters fuse with difficulty. Brown to green or greenish black. *Pyroxene* (Enstatite or Hypersthene), p. 43.
- *D.* Good cleavage in two directions at 90° or nearly so. Generally white, gray, flesh, or red. Scratched by quartz. Fusibility, 5. *Feldspars,* p. 25.
- *E.* In prismatic crystals, commonly twinned; scratches quartz; red brown to brownish black; intense ignition in closed tube gives a little water. *Staurolite,* p. 58.
- *F.* Reaction for *fluorine* when heated in tube with sodium metaphosphate. With cobalt nitrate reacts for aluminum (see *G* below). One good cleavage. Scratches quartz. *Topaz,* p. 63.
- *G.* Powdered mineral moistened with cobalt nitrate and intensely heated by the blowpipe on charcoal becomes blue (*aluminum*).
  - *a.* In stout rectangular prisms, commonly containing black carbonaceous impurities arranged as a cross or other symmetric patterns; not scratched by knife. *Andalusite,* p. 59.
  - *b.* In bladed, generally blue, crystals; scratched by knife parallel to long direction of best cleavage, but not at right angles to it. *Kyanite,* p. 60.
- *H.* No crystal form or structure. Effervesces in Na_2CO_3 bead. Yields a little water in closed tube on intense ignition. *Opal,* etc., p. 67.

PART III. ROCKS

CHAPTER VI

GENERAL PETROLOGY OF IGNEOUS ROCKS

All rocks, as already shown, fall into three great groups: igneous, sedimentary, and metamorphic. *Igneous* rocks are those formed by the solidification of molten masses derived from within the Earth.

Distinguishing characters of igneous rocks. The characters that distinguish igneous rocks from sedimentary and metamorphic rocks are of two kinds—field characters and specimen characters. The field characters are those that can be seen only in the field, by studying the rock mass in its relation to surrounding masses, or in other words by studying its geologic mode of occurrence in order to determine whether it is a dike, a laccolith, a lava sheet, etc. These igneous masses will be described presently. If the geologic mode of occurrence of a given rock mass can be positively ascertained, it indicates, more definitely than anything else, whether or not the rock mass is igneous in origin.

Commonly, however, a rock mass has boundaries that are so covered or obscured that its relations to the surrounding rocks and hence its geologic mode of occurrence cannot be determined. Moreover, it is often necessary to determine the nature of a specimen that was taken in the field from a rock mass and brought back for study. We are then compelled to make use of the inherent characters of the rock, which can be ascertained by examining the rock at its outcrop or in specimens obtained from outcrops. Three characters distinguish igneous from sedimentary and metamorphic rocks. They are:

a. Entire absence of fossils.

b. The material composition.

c. The arrangement of the material: its texture and structure.

The first character is an obvious one, but it is largely of negative value, because many sedimentary and most metamorphic rocks do not contain fossils.

The second character refers to whether the rock is made up wholly of glass, partly of glass and partly of mineral grains, or wholly of mineral grains; if it consists wholly of mineral grains, the kinds and relative amounts of the minerals are important. If a rock is made up wholly or partly of glass, it is certainly of igneous origin. The presence of certain index minerals is also proof of igneous origin, but such index minerals are few in number and there is no general rule by which a rock can be definitely ascertained from its mineral composition to be of igneous origin. The igneous rocks must be determined by a knowledge of the different kinds of igneous rocks themselves, as they are described in a later chapter.

As for the third character, many igneous rocks are massive and homogeneous in appearance: a surface of the rock in one direction is like a surface in any other direction, that is, igneous rocks do not as a rule have the stratified, banded, or foliated structures that are characteristic of sedimentary and metamorphic rocks. Other highly characteristic structures may be present in igneous rocks, such as the amygdaloidal; they are described later.

MODES OF OCCURRENCE OF IGNEOUS ROCKS

There are two chief modes of occurrence of igneous rocks: the *extrusive* and the *intrusive.* In the extrusive mode of occurrence the molten rock-matter (the magma) rising from the depths below has reached the Earth's surface, come out upon it, solidified, and formed a rock mass. In the intrusive mode of occurrence the magma stopped before it attained the surface, and it cooled and solidified surrounded by other rock masses of the Earth's upper crust. Each of these groups has several recognized subdivisions, depending, in the extrusive rocks, on the conditions under which the magma was ejected, and, in the intrusive rocks, on the structural relation that the mass bears to the rocks that envelop it. Following in our mind's eye the ascent of the magma from the depths, we shall describe first the intrusive and then the extrusive mode of occurrence.

Intrusive masses. Intrusive masses are the bodies termed *dikes, sills, laccoliths, necks, stocks,* and *batholiths.* Other intrusive masses have been recognized and described, but for simplicity's sake they can be regarded as modifications of those just mentioned. The simplest intrusive mass is a dike, which will be described first.

DIKES. A dike is the result of the filling of a fissure in rock masses by molten magma that was injected into the fissure rapidly, generally in a single pulse, and has solidified. It is tabular in shape, its extension in length and breadth being great compared with its thickness. It may "cut," that is, pass through, other igneous rocks, or it may cut sedimentary or metamorphic rocks, depending on the material traversed by the fissure. In passing through sedimentary rocks it cuts at an angle across the planes of bedding. If a tabular mass of igneous rock is parallel to the bedding, however, it is by custom termed a sill. Dikes range from those but a fraction of an inch thick up to those that are half a mile thick. Two or three feet up to twenty are the thicknesses most common. Dikes may extend but a yard or two long, as exposed on the surface, or they may be many miles long: a great dike in the north of England is more than a hundred miles long. The plane of extension of a dike is generally vertical or nearly so, but it may be inclined at varying angles to the horizontal plane. The angle of inclination is called the *dip* of the dike, and the direction its outcrop takes in intersecting the horizontal plane is the *strike* of the dike.

Dikes have attained the Earth's surface and given rise to outflows of lavas, or they may have become exposed only by subsequent erosion. During the course of erosion some dikes have resisted more strongly than the rocks that enclose them, and therefore they project as walls, a common feature. Other dikes composed of more easily weathering materials have resisted less well and have become ditches, which are less common. Dikes may show the columnar structure described later, the columns being at right angles to the walls. Dikes that have cut through sedimentary rocks have generally baked and altered them for some distance in the manner described under contact metamorphism. A dike that cuts a sheet of igneous rock enclosed in stratified beds is shown in Plate 2.

SILLS. It has frequently happened, when molten magma was being forced upward through, or into, stratified or other layered rocks, that it reached a place where the conditions were such that it could most easily spread out as a sheet between the strata or layered rocks. Intrusion of this kind has commonly taken place where the beds are weak and easily penetrated, as in shales and thinly bedded sandstones. The form of such an intrusive mass is like that of a dike, but unlike a dike it lies concordantly along the planes of stratification. Such intrusive sheets, or *sills* as they are technically called, vary from less than a foot to several hundred feet in thickness. They may have an areal extent of many miles; in fact the largest known sills have an extent of 5000 square miles. Like dikes, many sills have a columnar structure, the columns being perpendicular to the upper and lower surfaces and thus vertical. Some sills break, for limited distances, dikelike across the beds, to continue at a new level.

Sills are generally of the same massive, compact rock at top and bottom, and to some extent they have baked and altered the sedimentary beds both above and below them. These features distinguish sills from surface flows of lavas that were buried by sediments deposited upon them. Sills are most likely to occur in association with larger and more important intrusive masses, such as stocks and laccoliths, as accompanying or dependent features. In regions where

PLATE 2.

DIKE *D* (DARK COLORED), CUTTING AT RIGHT ANGLES ACROSS A SILL, *S* (LIGHT COLORED), AND ITS INCLOSING BEDS, *B*; CASTLE MOUNTAIN, MONTANA.

(U. S. Geological Survey.)

thick, extensive sills occur and the inclosing beds have been faulted, dislocated, and tilted, the sills give rise, as a result of erosion, to prominent topographic features, as illustrated by the trap ridges of southern New England, northern New Jersey, and many other places. Examples of sills are shown in Plates 2 and 3.

LACCOLITHS. Laccoliths are great lenticular masses of igneous rock lying between stratified beds that wrap around them. If, while forming a sill, magma is injected from below faster than it can spread laterally, it will begin to arch up the overlying strata, and if the magma continues to be injected a great, thick planoconvex lens of liquid rock matter is formed.* After solidifying, the resulting igneous mass constitutes a laccolith. A cross section of such a laccolith is shown in Fig. 61. Also, where the folding strata on the flanks of rising mountain ranges are subjected to horizontal pressure, they tend to open, and such openings may be filled with magma from below, *pari passu* with their opening, as illustrated in Fig. 62. In general, laccoliths are roughly circular or oval in ground plan. Although many are symmetric as in the diagrams, they are likely to be more or less irregular in shape. The floor may be flat or tilted, as in the figures.

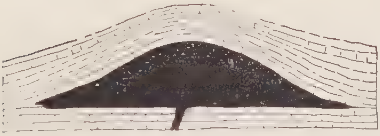

FIG. 61. Cross section of a laccolith.

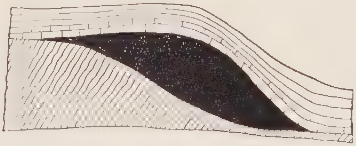

FIG. 62. Section of an inclined laccolith.

In many of the larger laccoliths the floor dips from the perimeter inward toward a point under the center of the laccolith, just as if the floor had sagged from loss of support. Such loss of support is conceivably due to the emptying of a magma reservoir below the laccolith. The immense Duluth gabbro mass of Minnesota, at least 2400 square miles in extent, is a notable example of a laccomorphic body having a centripetally dipping floor; it is called a *lopolith* (pronounced lō-pŏ-lith).

Laccoliths differ from sills only in being extremely thick in comparison with their lateral extension, and all gradations between them occur. A laccolith may be a mile or more in thickness and many miles in diameter, or but a few hundred yards across. The beds above are

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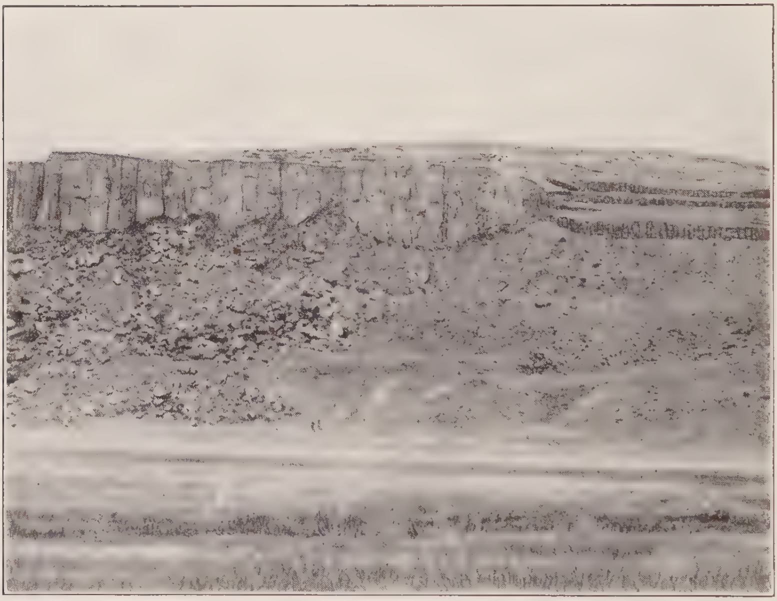
* Increasing viscosity of magma as the result of cooling during injection also helps to slow up the lateral movement of the magma and thus to arch up the overlying strata.

PLATE 3.



LACCOLITH AND FRINGING SILLS OF MELAPHYRE IN SANDSTONES; SHONKIN SAG, MONTANA.

(U. S. Geological Survey.)

usually stretched, thinned, and broken during the process of bending them to arch over the top of the laccolith. Like sills, laccoliths alter the strata above and below them by contact metamorphism. They occur most commonly in weak beds of shale, etc., the stronger, thicker beds of sandstone and limestone being up-arched. The best known examples are in the region of the Rocky Mountains where they are not uncommon. Plate 3 shows the Shonkin Sag laccolith and Plate 4 the nearby Square Butte laccolith, a striking landmark in that part of Montana.

Necks. When a volcano has ended its activity and become extinct, the column of magma that occupies the conduit leading to the depths below solidifies and forms a mass of igneous rock. In the course of time erosion cuts away most of the light porous ash and the lavas, leaving the more solid and resistant rock projecting, as shown by the heavy line *acb* in Fig. 63. The level of erosion may continue to descend into the rocks that form the foundation on which the volcano was built, and all traces of the ash and lava may eventually be swept away, leaving only the filling of the conduit to mark the former site of the volcano. Such a mass is termed a *volcanic neck*. Volcanic necks are more or less circular in ground plan and range from a few hundred yards up to a mile in diameter. The rocks surrounding them are generally cut by many radially arranged dikes, and if stratified they contain sills.

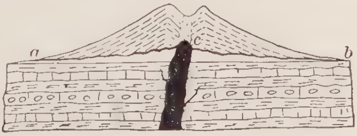


Fig. 63. Section through a volcano.

Stocks. Stocks are bodies of intrusive rock that ascended into the upper region of the Earth's crust in the form of magma and have solidified there. They have become exposed to view as a result of erosion. They are more or less circular or elliptic in ground plan. Their contact surfaces cut across the inclosing rocks and are more or less irregular, and most stocks become larger in plan in depth. Stocks range in areal extent from a few hundred square yards to many square miles. The distinction from a volcanic neck is not based on size, though necks tend to be smaller than stocks, but lies in the fact that the term neck is employed only when the evidence shows that there has been extrusive volcanic activity from the column of magma now represented by the rock mass. Some stocks were doubtless necks, but this cannot now be proved. The granite hills of New England, of Scotland, and of many other eroded mountain regions are mostly stocks or batholiths.

PLATE 4.

SQUARE BUTTE, A ZONED LACCOLITH WHOSE COVERING STRATA HAVE BEEN COMPLETELY STRIPPED OFF BY EROSION; HIGHWOOD MOUNTAINS, MONTANA.
(U. S. Geological Survey.)

Batholiths. The term batholith is widely used to designate huge intrusive masses of igneous rock that so far as known extend indefinitely downward. They are seen in deeply eroded areas of the Earth's crust, where they are characteristically accompanied or surrounded by metamorphic rocks, or in mountainous regions where they form the central cores of the ranges, as in the Sierra Nevada and in the Coast Range of British Columbia and Alaska. They differ from stocks only in their much greater size, some of them being many thousands of square miles in surface area. It has been suggested that intrusive masses less than 100 square kilometers (40 square miles) in areal extent be called stocks, and those larger than 100 square kilometers be called batholiths. Some authorities would place this limiting value, which is purely arbitrary, at 10 square miles. A large stock is a small batholith. The larger batholiths are of vast size: the Idaho batholith, occupying a large part of Idaho and lapping over into Montana, is 16,000 square miles in extent. The Coast Range batholith of British Columbia and southeastern Alaska is of the order of 100,000 square miles.

Although some stocks have clearly displaced the rocks whose site they occupy, the mode of emplacement of others and especially of batholiths is still a problem to be solved. Some hold that they have made room for themselves in the Earth's crust by melting and dissolving the rocks that formerly occupied the space now occupied by the igneous mass, whereas others urge the view that the invading magma actively made room for itself by crowding aside the surrounding rocks. Various modifications of these views have been suggested and even more remarkable ones have been put forward, but geologic science is not yet in a position to pronounce definitely as to their correctness.

Extrusive igneous rocks. Two chief modes of extrusion can be recognized: the *quiet* mode, giving rise to outwellings of fluent magma, which then solidifies to rock; and the *explosive*, in which the magma, by the violent action of gases, is ejected into the air and falls as innumerable fragments.

Quiet Eruption; Lava Flows. Magma that rises to the Earth's surface and flows out is called *lava*. The solidified material is a sheet of lava or *extrusive sheet*. Lava flows often issue from volcanoes, and the extrusions of some volcanoes, such as those now active in Hawaii, are chiefly of this nature. At other volcanoes they alternate with or succeed explosive ejections of fragmental material.

Some lava flows were not extruded at volcanic centers but were poured out as quiet outwellings from numerous fissures. Extrusion in this manner has occurred many times in the Earth's history on a huge scale, as in the Columbia River region of the northwestern United States and in western India. In these regions the resulting piles of superposed lava flows are thousands of feet thick and cover areas of more than 200,000 square miles.

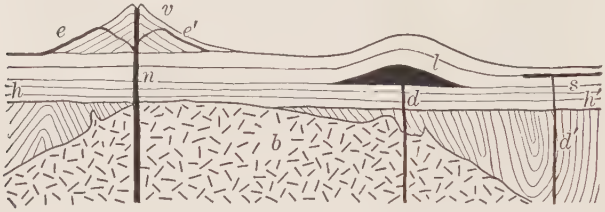


Fig. 64. Diagram illustrating the occurrence of igneous rocks: *b*, cross-cutting batholith intrusive into closely folded slates. The batholith was partly uncovered by erosion, and on this eroded surface (*hh′*) horizontal beds were laid down. Into these beds were injected a laccolith (*l*) and a sill (*s*), fed from dikes (*d*) and (*d′*); and on the Earth's surface was built the volcano (*v*) fed from the depths of the conduit (*n*); *e* is a lava flow and *e′* a lava flow buried by a later eruption of tuff. Horizontal distance shown, 30 miles; vertical distance, 3 miles.

Not infrequently sheets of lava have become depressed below sea level by downwarping of the Earth's crust and have been covered by deposits, or they were erupted on the sea floor and have been covered by contemporary sedimentation. Such buried extrusive sheets can be distinguished from intrusive sheets (sills) by the facts that they have not changed by contact metamorphism (page 149) the sedimentary rocks above them and that their tops generally show the structures typical of the surfaces of lavas, such as the vesicular, amygdaloidal, scoriaceous, and ropy surfaces described later.

Explosive Eruption; Tuffs and Breccias. When magma in the conduit of a volcano attains the surface, it either flows out quietly as already mentioned, or, if it is stiffly viscous and charged with gases under great tension, it becomes violently explosive. Clots of magma are blown into the air, finally to fall as splotches or as solid fragments, depending on how much they cooled during their flight. Owing to the expansion of the gases, chiefly steam, the ejected pieces are generally more or less vesicular. They range in size from those weighing several hundred pounds to dust so fine that it floats for long periods in the air. According to size, the ejected material is roughly classified as follows: pieces the size of an apple and upward are called *blocks* if angular and *bombs* if roundish or ellipsoidal; those the size of nuts are termed *cinders*; those the size of small peas or shot are called *volcanic ash*; and the finest is called *volcanic dust*. The coarser material—the bombs, blocks, and cinders—falls around the vent and builds up a cone; the lighter ash and dust, carried by air currents, tend to fall after the coarser material and at greater distances from the vent. The coarser deposit thus produced is termed *volcanic breccia*, and the finer is known as *tuff*.

GENERAL CHARACTERS OF IGNEOUS ROCKS

Since igneous rocks are formed by magmas solidifying, the nature of the resulting rocks must depend largely on the chemical composition of the magmas from which they were formed. Most igneous rocks are composed of mineral grains, and the kinds and relative amounts of these minerals necessarily depend on the kinds and relative amounts of the chemical elements that were in the magmas. It is pertinent, therefore, to examine the Earth's magmas in regard to their chemical composition and to see if any general rules govern their composition.

Chemical composition of magmas. We cannot, of course, subject a magma directly to investigation, but if an average sample of an igneous rock is subjected to chemical analysis, essentially the same result is achieved. Many thousands of such analyses have been made of rocks from all parts of the world. These analyses show that the rocks, and therefore the magmas from which they were derived, are made up essentially of a few chemical constituents. These constituents are reported by the chemist as oxides, as follows: *silica*, SiO_2; *alumina*, Al_2O_3; *iron oxides*, both *ferric*, Fe_2O_3, and *ferrous*, FeO; *magnesia*, MgO; *lime*, CaO; *soda*, Na_2O; and *potash*, K_2O. Other oxides, including water, are also reported, but as they occur in relatively small amounts they can be neglected. As will be shown later, however, the water and other volatile constituents are of great importance while in the magma.

The variation in the chemical composition of magmas is shown in a general way in Table 1. Although chemical analyses are by convention reported in terms of oxides, the oxides thus reported do not exist as such in the magma; in reality they are in the magma combined as mineral molecules, more or less electrolytically dissociated.

Variation of magmas and mineral composition. It is not to be understood that the analyses in Table 1 represent all the different varieties of magmas that there are; they were selected to show only the most prominent features of variation. Certain of these features are readily apparent. Thus in the first three analyses silica, alumina,

TABLE 1

ANALYSES OF IGNEOUS ROCKS

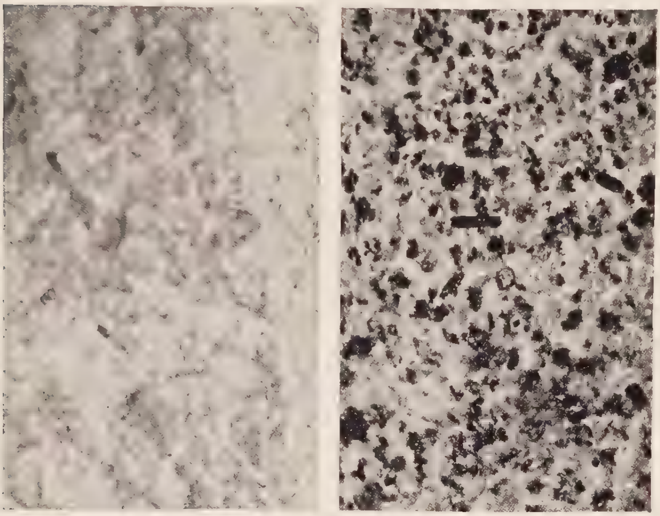
	I	II	III	IV	V	VI	VII	VIII
SiO_2	56.6	65.5	72.5	65.1	56.0	49.2	40.1	38.4
Al_2O_3	22.4	17.8	13.1	16.2	15.6	12.0	7.8	0.3
Fe_2O_3	1.8	0.7	1.7	1.1	1.2	2.8	7.3	3.4
FeO	0.8	1.2	1.0	3.2	6.3	8.8	8.6	6.7
MgO	1.3	1.0	0.6	2.3	6.8	9.3	23.7	45.2
CaO	0.3	1.9	1.0	4.0	7.3	10.6	6.5	0.4
Na_2O	8.5	5.6	4.2	4.0	2.2	1.9	1.2	0.1
K_2O	7.3	5.6	4.9	2.5	1.3	1.7	0.5	
Rest	1.4	0.7	0.7	1.6	3.3	3.2	4.5	5.7
Total	100.4	100.0	99.7	100.0	100.0	99.5	100.2	100.2

I. Nepheline syenite, Serra de Monchique, Portugal.
II. Syenite, Highwood Mountains, Montana.
III. Granite, Castle Mountains, Montana.
IV. Quartz diorite, Electric Peak, Yellowstone National Park.
V. Diorite, Montgomery County, Maryland.
VI. Gabbro, Red Mountain, Montana.
VII. Peridotite, Devonshire, England.
VIII. Dunite, Tulameen River, British Columbia.

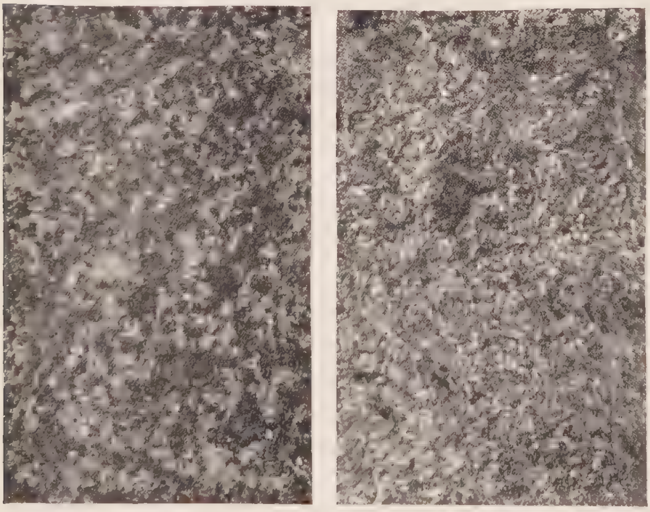
and the alkalies, soda and potash, are clearly the chief constituents, whereas lime, iron, and magnesia play minor parts. Therefore, when minerals crystallize from such magmas, they consist chiefly of alkali feldspars, because the alkali feldspars are composed of silica, alumina, and alkalies. Again, if we inspect the amounts of silica in these three analyses and recall that orthoclase, the potassium feldspar, contains 65 per cent silica, and albite, the sodium feldspar, about 68, it is clear that in specimen III there is more silica than needed to combine with the alkalies and alumina to form feldspars. Therefore silica will be left over after the feldspars have formed, and this extra silica will crystallize as free quartz. In I, on the contrary, there is not enough silica to turn all of the alumina and alkalies into feldspar, and a certain amount of some material, such as nepheline, which contains these oxides combined with a smaller amount of silica, must be formed to compensate for this deficiency. In II the amount of silica is very nearly that required to form pure feldspars, and these feldspars will therefore make up the bulk of the rock with little of either quartz or nepheline.

If we now turn our attention to the oxides of calcium, iron, and magnesium, manifestly the minerals that they produce, such as biotite,

PLATE 5.



A. Anorthosite, all feldspar. B. Syenite, mostly feldspar.



C. Diorite, some feldspar. D. Peridotite, no feldspar.

CONTRAST OF FELDSPATHIC AND FERRO-MAGNESIAN ROCKS.

hornblende, and pyroxene, will have but a subordinate role in the first three rocks. In IV to VII, however, these oxides increase and silica, alumina, and alkalies decrease, and eventually the last two vanish and the silica becomes very low. Clearly, in terms of minerals, in the five rocks formed from these magmas the *ferromagnesian* minerals—those containing iron or magnesia or generally both—such minerals as amphibole, pyroxene, or olivine, would play an increasingly important role. The last rock (VIII) would be largely or wholly composed of ferromagnesian minerals, and feldspars would be correspondingly less abundant or would be entirely absent. The contrasting appearance of the resulting rocks is illustrated in Plate 5.

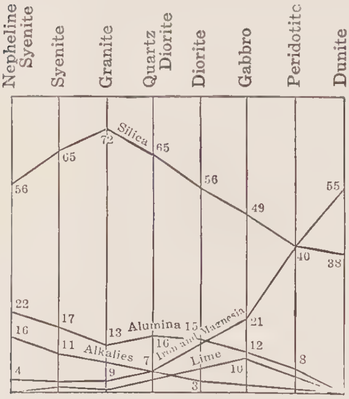


Fig. 65. Diagram to illustrate chemical variation of igneous rocks.

The variation in the amount of lime deserves a separate word, because lime has a dual function: it can form a feldspar with alumina and silica, which as a rule then combines with soda to form plagioclase; or it can enter into the ferromagnesian minerals, pyroxene and amphibole. It generally does both; and thus, as we follow the table of analyses from left to right, as lime increases, the quantity both of plagioclase and of ferromagnesian minerals increases also. Coincident with this the alumina also increases somewhat.

Variation shown by diagrams. The facts just stated can be shown graphically by means of a simple diagram, Fig. 65. Thus in the place of the analyses of the foregoing table we can draw vertical lines, one for each analysis, at equal distances apart, and on each line we can set off a vertical distance in millimeters equal to the per cents of each oxide in the analysis. Through these points lines are drawn corresponding to each oxide. Iron and magnesia, from the similarity of function they exhibit, are united in one line. The equal distances for each analysis at the foot of the diagram thus serve as abscissas and the percentages are ordinates, while the connecting lines approach curves that show the mutual relations of the oxides. When the variation in the amounts of the oxides was described, it was pointed out that this variation caused a corresponding variation in the minerals that crystallize from the magmas composed of these oxides. By considering the relative amounts of the important minerals that each type of analysis would produce, we can construct a diagram, Fig. 66, which shows in a general way the variation of the minerals in the common rocks. It also shows the relative proportions of the minerals in the more abundant kinds of igneous rocks, and it serves as a basis for their classification, as will be explained later. The relative proportions of the minerals are given in the vertical direction, and the variation and passage of one kind of rock into another in the horizontal direction.

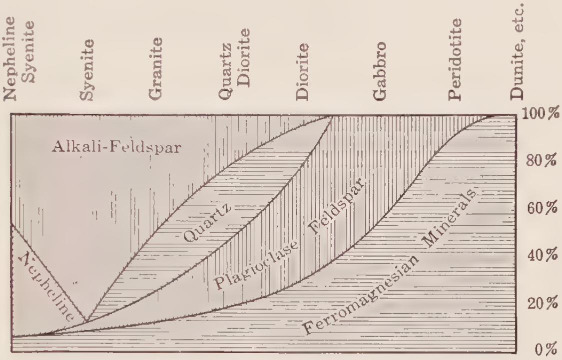


FIG. 66. Diagram to illustrate the variations and relative proportions of the minerals composing the important igneous rocks.

These diagrams and the table of analyses are not to be taken as rigidly representing all the possible mineral combinations of igneous rocks. They most certainly do not. Other analyses might be selected that would yield different diagrams, and if of rare and uncommon rocks they would be very different indeed. In a general way, however, these diagrams show the most important chemical and mineralogical features that distinguish the common kinds of igneous rocks from one another.

Minerals of igneous rocks. From what has just been said, manifestly the important minerals that compose the igneous rocks are feldspars, quartz, and the ferromagnesian minerals. For purposes of classification, to be explained later, the ferromagnesian minerals are contrasted with quartz and feldspars. Recalling that silica (SiO_2) and alumina (Al_2O_3) are prominent constituents of quartz and feldspars, we can call this group the *sial*ic one. The ferromagnesian minerals are called *mafic* minerals, or *mafites*, from *ma* in *ma*gnesian and *f* in *f*erro.

The prominent minerals of the igneous rocks are given in Table 2. The last mineral in the sialic group—nepheline—is much less important than the first three because of its restricted occurrence. The iron ore minerals—hematite, ilmenite, and magnetite (commonly called "iron ores" for short)—though so widely distributed that nearly all igneous rocks contain one or more of them, are less important than the other ferromagnesian minerals because as a rule they are only a minor part of the total bulk of the rocks. A mineral, such as these, which is evenly distributed through a rock but makes only a small part of its mass, is called an *accessory* mineral in contradistinction to those forming its main bulk, which are called *essential* minerals.

The chemical and physical characters of the minerals mentioned in Table 2 are described under the appropriate headings in Part II, which can be consulted for further information regarding them.

TABLE 2

CHIEF IGNEOUS ROCK-MAKING MINERALS

Sialic Group	*Ferromagnesian Group*
Alkali feldspar	Olivine
Plagioclase feldspar	Pyroxene
Quartz	Amphibole
	Biotite
Nepheline	
	Iron ores

Order of crystallization. If a polished surface of a coarse-grained rock is attentively studied with a lens, or better if a thin section is examined under the microscope, the minerals composing the rock will be seen to have differing degrees of perfection of their crystal forms. From this difference we conclude that the minerals have crystallized from the magma not simultaneously but successively. Thus in Fig. 67 grains and octahedrons of magnetite are inclosed in the crystals of biotite (M); they occur also in the other minerals. They are evidently older than the biotite, because they are inclosed by it. The biotite is older than the plagioclase feldspars (P), because it abuts into them with its own crystal faces or was partly inclosed by them as they grew around the already formed crystals. Similarly, the plagioclase has its own form where it adjoins the orthoclase (O) and the quartz (Q) which surround it, and it is therefore judged to be older than they are. When the orthoclase and quartz are examined, they are seen to shows no crystal boundaries with respect to one another, and they are therefore judged to have crystallized simultaneously or nearly simultaneously. The order of crystallization as thus determinable for this particular rock is: first, magnetite, then biotite, then plagioclase, and lastly orthoclase and quartz.

Igneous rocks show in general this order of crystallization: first, the oxides of iron, then ferromagnesian minerals, then plagioclase feldspars, then alkali feldspars (and feldspathoids), and lastly quartz. One sees from this, as illustrated in Table 3, that in the order of crystallization metallic oxides containing no silica are first, the ferromagnesian minerals (orthosilicates and metasilicates) next, feldspars containing more silica next, and quartz last. Thus minerals richer and richer in silica tend to crystallize out successively. It is not to be understood, however, that one mineral necessarily finishes its period of crystallization before another begins as in *A* in Table 4 but rather that, as in *B*, most of it has separated from the magma before the next mineral begins to crystallize.

TABLE 3

1. Magnetite, Fe_3O_4.
2. Pyroxene, $(Mg, Fe)Ca(SiO_3)_2$.
3. Plagioclase, Ab_xAn_y.
4. Orthoclase, $KAlSi_3O_8$.
5. Quartz, SiO_2.

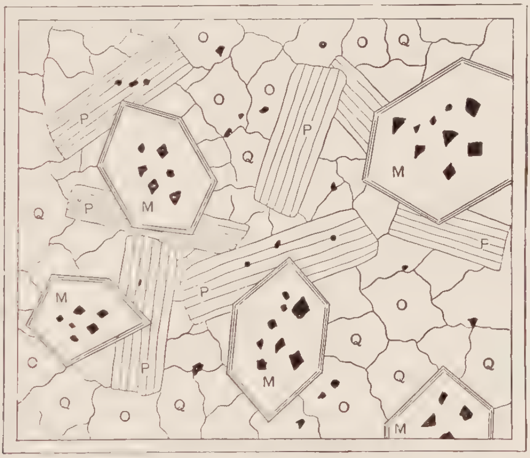


Fig. 67. Diagram to illustrate successive crystallization.

Insolubility vs. infusibility. A magma consisting of molten silicates is to be regarded as a complex solution of some compounds in others, like a solution of several salts in a solvent, such as water. As the hot solution cools, a point is reached where some compound, or mineral, become insoluble in the cooling solution and therefore begins to crystallize out. The statement sometimes made that the minerals crystallize in the order of their fusibility is essentially wrong; for in the preceding paragraph we saw that pyroxene crystallizes before before quartz, yet pyroxene fuses rather readily before the blowpipe, while quartz is infusible. Therefore the solubility, not the infusibility, determines the order of crystallization.

TABLE 4

A	Magnetite	Pyroxene	Plagioclase	Orthoclase	Quartz
B	Magnetite	....	....	....	....
	Pyroxene	....	....	....	....
			Plagioclase	....	....
				Orthoclase	....
					Quartz ....

Influence of mineralizers. Experience shows that magmas reaching the surface at volcanoes and as lava flows contain much volatile matter, especially water, which they give off, generally with explosive violence. Possibly magmas acquire part of their content of volatile matter from moisture-laden rocks with which they came in contact on their way to the surface. At present the volatile substances are thought to be chiefly or wholly of primal origin, that is, original constituents of the Earth's interior molten masses, contained therein under pressure. Without further regard to the theories advanced to explain the presence of the volatile substances, we know that magmas contain them and that they are of great importance in the formation of igneous rocks. Water is the most abundant of the volatile substances, but carbon dioxide, fluorine, boric acid, sulfur, and chlorine are also prominent and may produce important results.

The work of various investigators, especially French, has shown that some minerals, such as pyroxene, magnetite, calcic feldspar, olivine, and nepheline, can be made artifically by fusing their constituents together and allowing the molten mass to cool slowly. Other minerals, such as hornblende and biotite, do not form in "dry" fusions, however. To produce hornblende and biotite volatile substances, especially water, must be present. These substances act in three ways: one, chemically, since some minerals, such as biotite and hornblende, contain small quantities of water (in the form of hydroxyl, OH), fluorine, or both, and these substances are consequently necessary in forming hornblende and biotite; two, physically, in that these substances lower the melting point of the magma; and three, in greatly increasing the fluidity of the magma. Thus orthoclase, albite, and quartz, which have high melting points, in a dry fusion are extraordinarily viscous; and on cooling they are therefore unable to crystallize and consequently solidify as glasses. The presence of water under pressure lowers the temperature of solidification and decreases the viscosity of the molten mass; the internal friction being thus diminished, the ions can move freely and can arrange themselves in crystal form, thus producing orthoclase, albite, and quartz. As crystallization progresses, the volatile substances, except the amount chemically (and to a minor extent mechanically) retained in the minerals, are excluded from the solidifying rock mass to play an active role in new and important processes, such as forming pegmatites, causing contact metamorphism; other such processes are described later in their appropriate places.

TEXTURE OF IGNEOUS ROCKS

It has been pointed out that igneous rocks are diverse in the kinds and proportions of their component minerals and that this diversity is mainly determined by the chemical composition of the magmas from which they were derived. Another important way in which igneous rocks differ is in their texture. By texture is meant the appearance of a rock as determined by the size, shape, and arrangement of its constituents. Thus some rocks are made up of grains so large that the identity of the minerals comprising the grains is easily determined, whereas other rocks are made up of grains so small that they defy identification by the unaided eye or simple lens. Again, the constituent grains of a rock may all be of about one size, or they may range in size, some being relatively large and the rest minute. Furthermore, the magma may have solidified as a glass, or as a mixture of glass and crystals. Such variations are independent of chemical composition: they depend on the physical conditions under which the magma has solidified. Thus one and the same magma of a given chemical composition will yield rocks of diverse textures under diverse physical conditions.

Factors determining texture. If a strong hot solution of a salt, such as common alum in water, is allowed to cool slowly and continuously, few centers of crystallization are spontaneously set up, and the resulting few crystals, having a long period in which to grow, will be of good size. If, on the contrary, the solution cools rapidly, many centers of crystallization form spontaneously and a large number of crystals begin to grow, every one of which, sharing its nourishment with a large number of associates, remains very small. The same holds true for the molten liquids from which igneous rocks are formed. If the magma cools slowly, coarse-grained rocks are produced; if it cools rapidly, fine-grained rocks are formed; and if the cooling takes place so quickly that there is no opportunity for crystallization even to begin, rocks wholly or in part composed of glass will result. The *rate of cooling* then is a prominent factor in determining the texture of rocks from a given magma.

The direct effect of pressures prevailing in the upper levels of the Earth's crust is not a prominent factor in determining texture, for geological observation has shown that a particular variety of texture extends unchanged through a range of 10,000 feet vertically. The effect of pressure is indirect, by causing the magma to retain its volatile matter until a late stage of solidification.

The *presence of volatile matter*, especially water, has a strong influence in determining the texture, particularly in the silicic rocks, for the volatile matter greatly increases the fluidity of magma. As the magmas cool down and the crystallizing points of the different minerals are reached, they retain their fluidity* if they retain their content of volatile matter, instead of becoming highly viscous. This fluidity allows the ions to move easily as they migrate toward the growing crystals and enables these crystals to grow large; therefore, a coarse-grained texture is produced.

The *chemical composition* of magmas has an important influence in producing texture. This influence shows itself in two ways. Magmas deficient in silica, especially those containing much iron and magnesia (shown in the right-hand side of Fig. 65, page 116), remain liquid without becoming stiffly viscous at much lower temperatures than those high in silica (shown in the middle of Fig. 65). This fluidity of the magma enables the minerals to crystallize more freely and in consequence to form coarse-textured rocks under circumstances in which silicic magmas, because of their high viscosity, yield only fine-grained rocks or even glassy rocks. The other way in which chemical composition influences texture is as follows. Differences in composition in the magmas mean, of course, differences in the kinds of minerals that they produce. Different minerals crystallize in their own distinctive shapes, and although, owing to interference with one another, they may not succeed in forming perfect crystals, they tend to approximate such shapes. Some are tabular, others are spherical or cuboidal grains, and others are elongate prisms. Thus, although the general size of such grains may remain the same throughout a mass of rock, the differences in the shapes of the grains produce differences in the pattern or *fabric* of the rock and thus influence its texture.

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* Fluidity is the inverse of viscosity.

Relation of texture to geologic mode of occurrence. The conditions most favorable for producing coarse-grained rocks—slow cooling and retention of the volatile constituents in the magma—will, in general, be best realized when the magma comprises a large mass that solidifies within the Earth's crust, so that it is completely enveloped by the surrounding rocks. The rock cover acts as insulator and causes the magma to retain its heat and volatile matter. The resulting igneous rocks, formed in depth, have become exposed to our observation only after erosion has carried away the enveloping rock cover. They are commonly called *plutonic*, *deep-seated*, or *abyssal* rocks—terms it would be best to abandon because of the exaggerated ideas of depth that they convey. From what has been said of the modes of occurrence of the igneous rocks, batholiths, stocks, and the lower portions of volcanic necks can be expected to be coarse grained, and as a matter of fact they nearly all are.

On the other hand, when magma reaches the Earth's surface and pours out in volcanic eruptions and lava flows, entirely different conditions prevail: no insulating cover retains the heat and consequently the magma cools rapidly. Also, the pressure has been relieved, and with loss of cover and pressure the volatile matter quickly escapes. As a result, fine-grained rocks are formed. The cooling may indeed be so rapid that little or no crystallization takes place, and consequently rocks that are entirely or partly composed of glass are produced. If rocks are more or less glassy, it is presumptive evidence, though not proof, that they solidified as surface lavas.

In the minor intrusive bodies, such as dikes, sills, and smaller laccoliths, the conditions in general are intermediate between the two sets just described. The volume of magmas relative to the surrounding rocks was smaller; consequently heat and mineralizers were more rapidly lost than in the stocks and batholiths, and, since the depth in general was less, the pressure was less. Accordingly, the textures are generally intermediate between those of the larger plutonic mass and those of the lavas. But the physical conditions in these masses were very diverse, and in accord with this we find that the textures in some of them are aphanitic like those of the lavas—but rarely glassy—and some are coarsely granular like those of the large plutonic masses. Moreover, in some of the smaller intrusive bodies the influence of chemical composition in determining grain size described in the preceding section is conspicuously shown. Thus highly silicic dikes and sills, which are fine grained, are found to be associated under the same geologic conditions with basic dikes and sills, which are much coarser in grain.

Textures of igneous rocks. There follows a classification of the megascopic textures of igneous rocks, based on the principles outlined in the foregoing sections.

Phanerocrystalline: the rock consists of grains large enough to be recognized as individual entities by the unaided eye. *Example,* ordinary granite.

Microcrystalline: the rock consists of grains that can be perceived but are too small to be recognized.

Aphanitic: the rock consists of grains that are too minute to be perceived by the unaided eye. *Example,* many felsites.

Glassy: the rock can be distinctly seen to be wholly or predominantly composed of glass. *Example,* obsidian.

The distinctions stated above refer in part to the *crystallinity,* or proportion of crystalline material in the rock, for there exists a continuous series of rocks, ranging from those composed wholly of glass, through those consisting partly of glass and partly of crystals, to those made up wholly of crystals. In order for glass to be perceived by the unaided eye, however, it must make up most of the rock. The distinctions refer also in part to the absolute sizes of the crystal grains, to what is called the *grain size.*

Phanerocrystalline, or phaneric (Greek, *phaneros,* visible, manifest, evident), rocks are divided according to the size of grain as follows (see parts 1, 2, and 3 of Plate 6).

Fine-grained: the grains are less than 1 millimeter in diameter.

Medium-grained: the grains are between 1 and 5 millimeters in diameter.

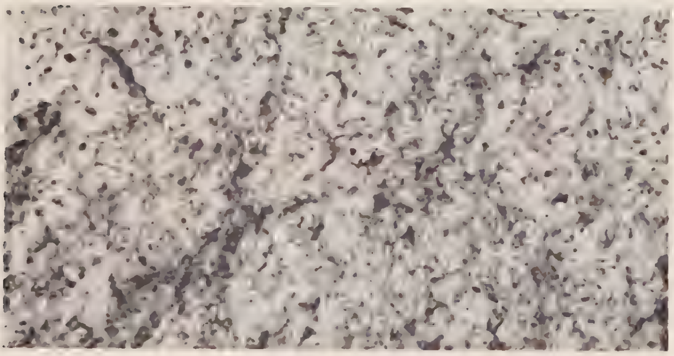
Coarse-grained: the grains are greater than 5 millimeters in diameter, or as large as or larger than peas.

Another important feature of texture is the pattern, or *fabric.* In megascopic work, the fabric is chiefly due to the relative sizes of the crystal grains in a given rock. Two chief kinds of fabric can be distinguished: *equigranular* fabric, in which all grains are roughly of the same general size, and *porphyritic* fabric, in which some grains are conspicuously larger and therefore contrast with the smaller ones or with the glass in which they are inclosed.

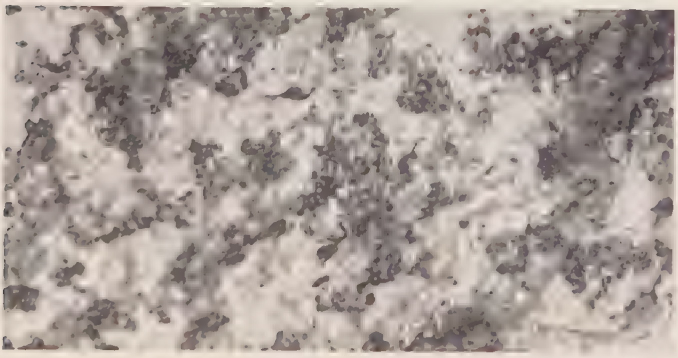
EQUIGRANULAR FABRIC. In an equigranular rock the crystal grains have roughly the same general size, as is shown in Plate 6. They do not, however, have necessarily the same shape. Careful examination of granites, which typically have the equigranular fabric, will show that the dark mica is generally present in well-formed (euhedral) hexagonal tablets or crystals, whereas the feldspars and quartz are in irregular (anhedral) masses, or the feldspar tends to have rough

PLATE 6.

1. Fine Grain.



2. Medium Grain.



3. Coarse Grain.

EQUIGRANULAR TEXTURE.

tabular or brick-like shapes. These differences in the attainment of their own crystal forms by the component minerals depend on the specific power of the minerals to develop their own characteristic crystal forms in competition with adjacent growing crystals and on the order of crystallization as previously explained.

PORPHYRITIC FABRIC. PORPHYRY. The porphyritic fabric, in its typical development, is characterized by a conspicuous contrast between large euhedral crystals, called *phenocrysts,* and the finer grained material in which they are embedded, called the *groundmass.* This groundmass may have any of the textural characters described on a preceding page: it may be equigranular, coarse or fine; it may be aphanitic or wholly or partly glassy. A rock consisting of phenocrysts embedded in a groundmass is said to be *porphyritic.** Examples are shown in Plate 7.

The phenocrysts may be scarce, abundant, or extremely abundant with respect to the groundmass. They may range in size from an inch or more in diameter down to dimensions so small as to require close observation to detect them. They may be light-colored feldspars and quartz or dark-colored ferromagnesian minerals (biotite, hornblende, pyroxene, and olivine), or minerals of both kinds. Furthermore, many phenocrysts are extremely well shaped and afford such striking specimens of perfectly formed crystals that they find a place in mineral cabinets, but others are poorly defined in crystal form. With increase in number of phenocrysts in a rock, and consequent crowding resulting in poor crystal form, all degrees of transition from porphyritic fabric into the equigranular fabric can be found. Porphyritic fabric is extremely common in lavas and in intrusives of small volume, such as dikes, sills, and small laccoliths; it is rare in the plutonic rocks, but occurs in some, especially in granites.

ORIGIN OF PORPHYRITIC FABRIC. It is easy to understand why many lavas are porphyritic. The lavas as they issue at the Earth's surface are full of growing crystals, many of considerable size, suspended in the molten fluid. The magma as it flows out becomes subjected to new conditions and is forced to cool rapidly; consequently it solidifies as a fine-grained, minutely crystalline, or even glassy, solid mass in which the earlier formed large crystals are

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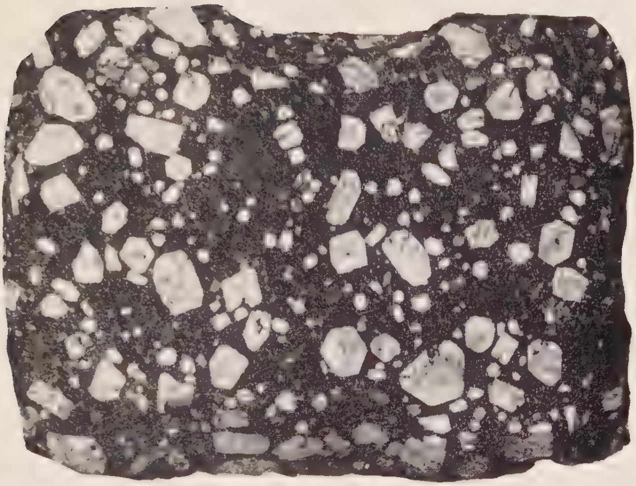
* The porphyritic fabric is not a contrast of *colors* but of *sizes* of mineral grains. Care must therefore be taken not to consider as a porphyry, for instance, a white rock that consists chiefly of grains of light-colored minerals such as feldspar, among which are imbedded a few conspicuous black grains of a ferromagnesian mineral of the same size, such as hornblende.

PLATE 7.



A. With Phenocrysts of Feldspar.

B. With Phenocrysts of Augite.

PORPHYRITIC FABRIC.

embedded. Thus the completed rock has a porphyritic fabric. An essentially similar process serves to explain the development of this fabric in some of the smaller intrusives, such as dikes and sills. It cannot serve as a general explanation, however, because in some dikes, sills, and laccoliths there is good evidence that the phenocrysts were not brought from afar during the injection of the magmas but were formed, like the rest of the rock, at the place where we now see them. It fails also to explain the porphyritic border of many granite masses and the large phenocrysts of other granites; nor does it explain the origin of the phenocrysts themselves, or explain why a few large crystals have formed while the rest of the magma fails to crystallize. Evidently a more general explanation is needed.

It has been previously shown that magmas are solutions of certain compounds in others. Therefore, as a magma cools, it becomes saturated. Now, some saturated solutions cannot crystallize spontaneously but require to be inoculated with a minute fragment of the substance in solution; they are in what is called the *metastable* state. Other saturated or supersaturated solutions either crystallize spontaneously or can be induced to crystallize by shaking or stirring with a foreign substance; they are in the *labile* state.

Miers has shown that the same solution may pass from one to the other of these states in accordance with changes of temperature, and suggests that a magma in the metastable state, in which a relatively few crystals induced by inoculation from the surrounding rocks are growing as phenocrysts, may by cooling pass into the labile state, whereupon the remaining liquid will spontaneously crystallize and form the groundmass. Or the magma may start in the labile state and be reduced by the formation of a crop of phenocrysts to the metastable state, in which condition it may be erupted as a lava; or, remaining and cooling down, it may pass into a new labile state, crystallize, and form the groundmass. The recognition of these states in cooling saturated solutions, like molten magmas, seems quite sufficient to explain the different variations of porphyritic texture that occur.

SOME STRUCTURES OF IGNEOUS ROCKS

The term *texture,* as already explained, is reserved for those features of the rocks that are produced by the size, shape, and arrangement of the component crystal grains. Certain larger features shown by the rocks are classed under the term *structure* and will now be described.*

Vesicular structures. When a magma rises to the Earth's surface, and especially when it issues as a lava flow, the pressure upon it is relieved and the water and other volatile matter it contains are given off. The escaping vapors and gases have a tendency, if the magma is still soft and stiffening, to puff it up into spongy vesicular forms, as illustrated in Plate 8. Silicic lavas may, indeed, be entirely changed into a light glass froth called *pumice*. Such frothed rocks are produced especially in the lava in the throat of a volcano, where the gases escape rapidly, or in the top portion of a flow. Except to a minor extent in dikes and sills, vesicles do not occur in intrusive rocks, and the presence of well-marked vesicular structure can be taken as a fairly strong indication but not proof that the rock in which it occurs was originally a lava flow. From the throat of a volcano vesicular lava may be blown out as fragments by explosions of steam, to fall as dust, ash, and cinders and form volcanic tuffs and breccias.

Amygdaloidal structure. Amygdaloid. After a lava has become vesicular, it may be permeated by hot waters that carry dissolved material, and the dissolved material may be deposited as minerals in the vesicles. Such deposition has happened notably in basaltic lavas, and the dark basalt then appears to be filled with round or ovoid bodies which, from the resemblance in shape of many of them to almonds are termed *amygdules*, from the Greek word for almond. The structure is called *amygdaloidal*, and a rock having this structure is an *amygdaloid*. An amygdaloid is shown in Plate 8. Although the smaller cavities are usually filled solidly, the larger ones may be hollow but lined with minerals projecting in crystals from the walls, as in geodes. From such amygdaloidal cavities some of the most beautiful crystal specimens are obtained. The minerals most prevalent are quartz, some of which is of the amethyst variety, calcite, chlorite, and particularly the zeolites. Basaltic lavas are the especial home of the zeolites, the more common kinds being analcite, stilbite, natrolite, heulandite, and chabazite.

The amygdaloidal structure is most widely prevalent in lavas, that is, in extrusive rocks, but it occurs also in some intrusive masses such as dikes and sills, especially at their margins.

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* The difference between the two usages can be illustrated as follows. A lava sheet might as the result of flowage appear to be made up of layers; the layers consist of rock composed of exceedingly fine particles. We say then that the lava has a layered structure and a very fine-grained texture.

PLATE 8.

A. VESICULAR LAVA.

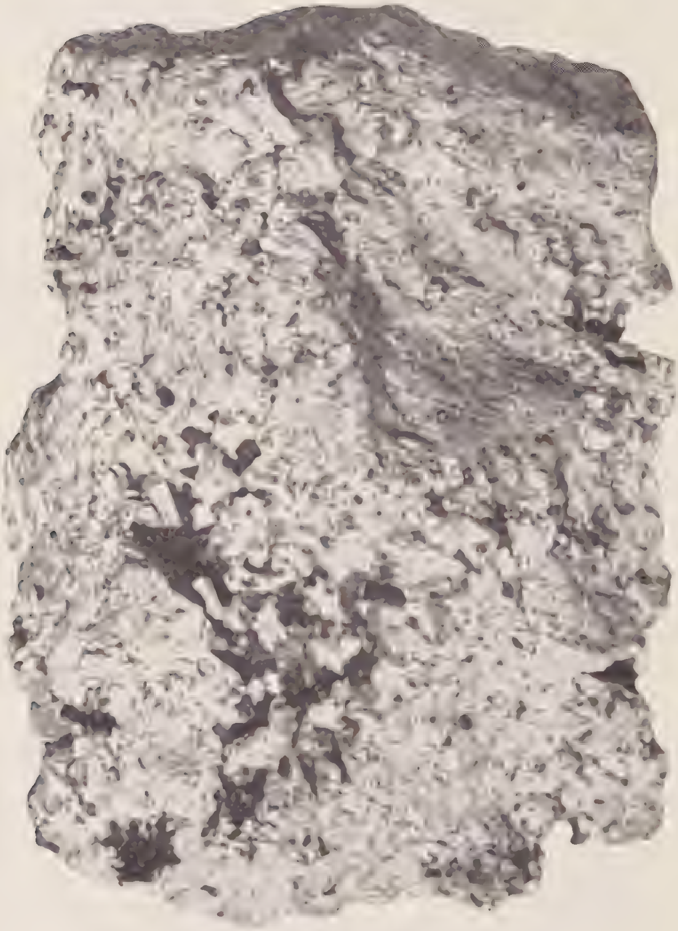
B. AMYGDALOIDAL BASALT.

Miarolitic structure and porosity. The volume of a magma is considerably greater than that of the solid crystalline rock formed from it. The volume of the magma is somewhat greater than when cooled to a glass. The decrease in volume in passing from the glassy state into the crystalline state causes a considerable rise in specific gravity. Thus an obsidian glass, consisting chiefly of silica and moderate amounts of alkalies and alumina, has a specific gravity of 2.3 to 2.4, but the same material crystallized into a quartz-feldspar rock (granite) has a specific gravity of 2.6. A notable reduction in volume has taken place, amounting to about 10 per cent.

In general, the decrease in volume during the process of crystallizing and the subsequent cooling produces minute interspaces or pores between the mineral grains, and cracking and jointing of the rock mass, a feature described in the following section. The pores formed account for the capacity of the rocks to absorb moisture. The porosity appears to be greatest in the coarse-grained rocks, much less in the finer-grained ones; greater in granites, less in diorites and other ferromagnesian rocks. In vesicular lavas the amount of pore space may be very great, but in ordinary crystalline igneous rocks it is small, as a rule less than 1 per cent of the rock volume.

In some rock masses, however, distinct cavities have formed during solidification. They are commonly small, usually an inch or so in diameter and rarely as much as several feet. The grain size of the rock bounding the cavities is much larger than that of the rock farther away, and the minerals of the boundary rock project into the cavity with free ends, bounded by distinct faces and having good crystal form. Also, especially in granites, the quartz and feldspar crystals are generally accompanied by those of muscovite, topaz, tourmaline, and other minerals foreign to the general mass of the rock but common in pegmatite veins. They also are euhedral. The presence of the water, fluorine, and boron necessary to produce these minerals, together with the larger size and distinct form of the crystals, shows that such volatile components, excluded elsewhere from the magma during the process of crystallization, collected in these cavities, possibly helped to enlarge them, and promoted the formation of the unusual minerals and the good crystal forms that they and the ordinary rock minerals exhibit. Such hollow spaces are called *miarolitic cavities*, and a rock containing them is said to have a *miarolitic structure*. Such crystal-lined cavities, or miaroles, can be distinguished from geodes and other cavities, in which the minerals were deposited long after the inclosing rock had been formed, by the fact that they have no distinct wall that separates the

PLATE 9.



MIAROLITIC CAVITIES IN GRANITE.

minerals from the surrounding rock. Miaroles are likely to contain fine mineral specimens. A miarolitic cavity is shown in Plate 9.

Jointing of igneous rocks. After a body of magma has solidified into rock and cools down to the temperature of its environment, the most important visible effect of the contraction that inevitably takes place is the presence of joints. These joints are the cracks or fissures which, running in various directions, divide the mass into blocks fitted together like masonry, as a rule according to more or less definite systems. The jointing may show itself in the formation of rudely cubic or rhomboidal blocks, as seen in granites and other related rocks, or as a platy parting which may be quite thin and cause the rock mass at first glance to resemble sedimentary beds. Platy partings and jointing on a small scale by which the rock body is divided into little blocks are most common in small intrusive masses—dikes and sills—and in lavas.

Jointing is a matter of great geologic importance because it permits air and water to enter into the body of the rock mass and accelerate the progress of weathering. As can be readily inferred, it is also of great practical importance in the work of rock excavation, in mining operations, and in quarrying. (See Plate 10.) Were it not for such joints, almost every igneous rock mass could furnish suitable material for quarrying; actually it is difficult to find a granite mass in which the joints are so widely spaced that it will furnish single blocks of great size, for example, those from which the celebrated Egyptian obelisks were made.

Columnar structure. The jointing of a cooling mass of igneous rock manifests itself most remarkably in the production of columnar structure. This structure is seen both in intrusive and extrusive masses and in all kinds of igneous rocks. It is most commonly displayed in basalts. The whole mass is made up of columns, regularly fitted together, from a few inches to several feet in diameter and from one foot to five hundred feet in length. A striking example from the High Sierra is shown in Plate 11. The celebrated Giant's Causeway on the north coast of Ireland is one of the best known examples. In perfect examples, as in the Giant's Causeway, the cross sections of the columns are mainly regular hexagons, and the columns are divided lengthwise at longer or shorter intervals by cross joints. The columns are perpendicular to the fastest cooling surface of the igneous mass; hence in a lava flow or in a sill they are vertical—on the assumption that the flow and sill are horizontal—whereas in a dike they tend to be horizontal. Such a dike when exposed by erosion tends to resemble a stretch of cord-wood regularly piled.

PLATE 10.

A. High Isle Quarry, Maine.

B. Allen Quarry, Mount Desert, Maine.

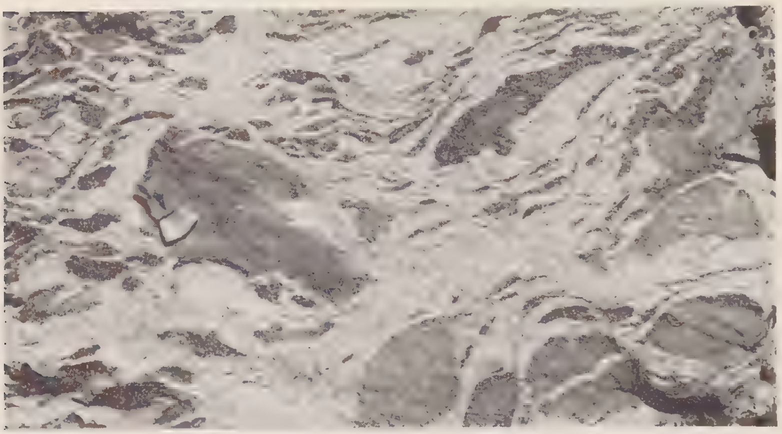
JOINTING IN GRANITE AND ITS USE IN QUARRYING.
(T. N. Dale, U. S. Geological Survey.)

PLATE 11.

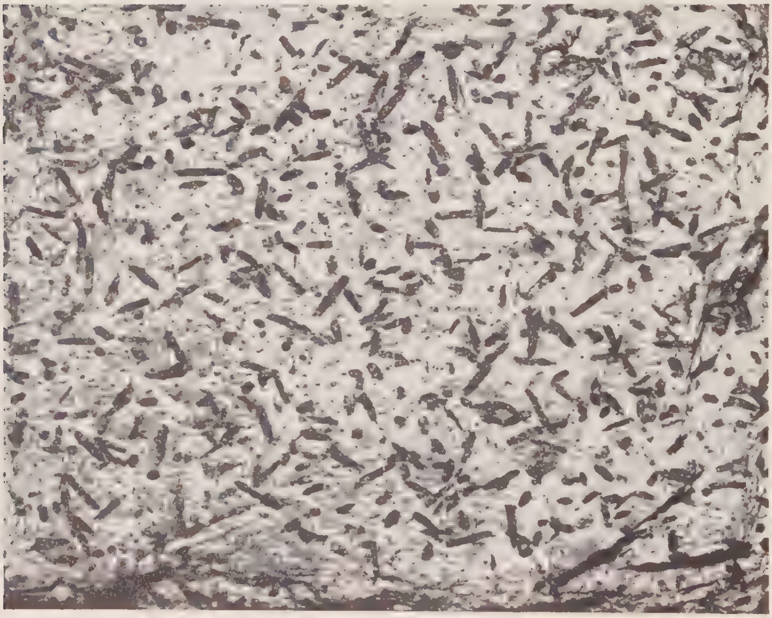
COLUMNAR STRUCTURE IN BASALTIC LAVA, SAN JOAQUIN RIVER, CALIFORNIA.

(U. S. Geological Survey.)

PLATE 12.



A. XENOLITHS IN GRANITE.
(F. C. Calkins, U. S. Geological Survey.)



B. CONTACT-METAMORPHIC SLATE, SAXONY.
Slate altered by contact metamorphism. Black crystals are cordierite altered to biotite.

The cause of columnar structure appears to be as follows. When a homogeneous mass is cooling slowly and regularly, centers of cracking tend to originate on the cooling surfaces at equally spaced intervals. From each center three cracks radiate outward at angles of 120° from each other. These cracks on intersecting produce regular hexagons and, penetrating downward, make columns. This regular arrangement produces the greatest amount of contraction with the least amount of cracking, provided that the centers are equally spaced. But as the contractional centers are not all equally spaced, three-, four-, five-, seven-, and eight-sided columns are formed. The columns themselves, contracting lengthwise, break into sections as they form. The same principle, *i.e.*, maximum contraction with minimum cracking, is exemplified in drying mud-flats, which crack into polygonal shapes, and in the prisms of drying and contracting starch. The well-developed columns in igneous rocks, however regular their appearance, are not crystals but pieces of rock and should not be confused with the hexagonal prisms produced by the crystallization of certain minerals, such as quartz and beryl, which are formed by an entirely different process.

INCLUSIONS IN IGNEOUS ROCKS

Frequently igneous masses contain bodies that differ in mineral composition, color, and texture from the inclosing rock. The inclosures commonly range in diameter from a fraction of an inch to several yards, but immensely larger ones may occur. Examination shows that some of them present all the characters of distinct kinds of rocks, and by these, and by their angular shapes, they show clearly that they are either fragments of pre-existent rocks that the magma on its way upward has torn loose from the walls of its conduit and brought along, or else they are blocks from the roof or sides of the chamber in which the magma came to rest, which were spalled off by the magma and sank into it. Inclusions of this mode of origin are called *xenoliths* (stranger rocks). They are composed of other kinds of igneous rocks or of sedimentary ones, such as shales or limestones. Inclusions of igneous rock are not commonly much changed, but the fragments of sedimentary rocks such as shale and limestone generally show the results of intense metamorphic action as described elsewhere, and accordingly are much altered. Xenoliths in the large intrusive masses occur most commonly near the borders. Xenoliths in granite are shown in Plate 12A.

Other inclusions are composed of certain minerals, especially ferromagnesian minerals, which have been aggregated by some process into lumps or nodules, such as the masses of olivine crystals that are inclosed in some basalts. The assembling of these minerals has possibly taken place while the remainder of the rock was still in a liquid condition; if so, the resulting masses are called *segregations.* Many xenoliths and segregations cannot be readily told apart; "inclusions" is, therefore a safer, noncommital term.

Some inclusions are indefinite in form and have blurred boundaries; they appear to have been drawn out, are lenticular, and are streaky in character. They may consist of the same minerals as those making up the rock in which they are inclosed but in quite different proportions, or they may contain different minerals. Thus dark streaks occurring in granite are much richer in hornblende or biotite than the inclosing rock. Some of these streaks are xenoliths that may have been melted up or otherwise altered by the magma, and most of them have probably been formed in this way. Some are possibly segregations that have been smeared out into streaks by movements of the inclosing magma. Such streaky portions are called by the Germans *schlieren,* and in default of anything better this word is often used for them in English.

Some lavas show a streaky or even a well-banded structure, consisting of portions that differ from one another in composition or in texture as the result of being drawn out during the flowage. This structure is called *eutaxitic.* Some remarkable eutaxitic rocks, however, are the results of the compaction, welding, and flowage of hot tuff and breccia deposited by pelean clouds, as at Crater Lake National Park.

ORIGIN OF IGNEOUS ROCKS—DIFFERENTIATION

The fact that lavas differing decidedly from each other in mineral and consequently in chemical composition have been erupted by the same volcano at different times early attracted the attention of geologists and led to much speculation as to its cause. Thus felsites and basalts frequently occur as the products erupted from a single vent. One explanation formerly used was that within the Earth there are two layers of magma: an upper one rich in silica, alumina, and alkalies; and a lower one, poor in silica but rich in iron and magnesia. According as the erupted magma was derived from one or the other of these layers, felsites or basalts were produced, and commingling of the two magmas in variable proportions gave rise to the rocks intermediate between felsite and basalt. However, both on chemical and geologic grounds, this view was soon seen to be insufficient to explain the origin of all igneous rocks.

As the study of rocks progressed, other facts of a similar nature came to light. Thus the mass composing the neck of an old volcano,* where the conditions of cooling were favorable for producing rocks of an equigranular texture, may be found to be composed of rocks of two or more distinct kinds. The boundary between the two shows in places that one was erupted after the other had solidified, since fragments of the one are inclosed in the other. This association is, of course, merely the downward extension into the volcanic conduit of the same relationship shown at the surface by the felsite and basalt lavas. Other [illegible] are of different nature and are of such geologic importance that they demand separate consideration.

Border zones. [illegible] masses, such as stocks and batholiths of granite or syenite, that have become exposed by erosion have an outer border of rock that differs in mineral composition from that of the interior of the mass. The rock of the border zone is connected with that of the interior by continuous gradation; it is said therefore to be a *facies* of the interior, or main mass. The thickness of such a border zone is variable, even in the same mass, and it may in places be lacking; it may be several thousand feet thick. Although in general it bears some proportion to the size of the whole mass, *i.e.*, the borders are larger in batholiths than in stocks, there is no definite rule.

A border facies as a rule is richer in the ferromagnesian minerals, such as pyroxene, hornblende, biotite, and magnetite, than the main mass into which it grades. The enriching minerals are generally the same as those occurring more sparsely in the main rock body, but different ones may be present. Consequently the border zone is richer in iron and magnesia, and to some extent in lime, than the main mass, and [illegible] in silica, alumina, and alkalies. Since they contain less of silica and hence are more basic, border zones of this kind are commonly called *basic border zones.* Not all border zones, however, are basic; [illegible] poorer in ferromagnesian minerals than the interior of the rock body and therefore contain more silica, alumina, and alkalies, a condition that expresses itself mineralogically in a greater [illegible] or of feldspar and quartz. They are accordingly called *acidic border zones.* As examples of basic border zones we have intrusive masses of syenite passing at their borders into pyroxenite; and as examples of acidic border zones we have syenite intrusive masses that become granite in their peripheral portions. As already pointed out, there is no definite contact between the rock of the main mass and that of its border zone; the one kind passes by continuous gradation into the other without change in texture. All

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* See volcanic necks, page 109.

the facts indicate that this arrangement was not produced by successive intrusions of different magmas, but by some process acting within the body of magma after it had arrived in place.

Zoned laccoliths. The zonal arrangement just mentioned is still more strikingly shown in certain laccoliths in Montana and elsewhere. Where these laccoliths have been laid bare and dissected by erosion, they are seen to consist of an exterior body of rock, composed largely of pyroxene, inclosing a core of rock of a totally different kind, generally a syenite, which is of course chiefly feldspar. A cross section through such a laccolith is shown in Fig. 68.

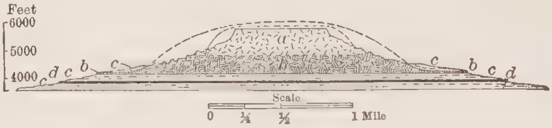

Fig. 68. Diagram of a zoned laccolith, Square Butte, Montana: *a*, feldspar rock; *b*, pyroxene rock; *c*, shales and sandstone; *d*, underlying sill of melaphyre. Figures in feet are heights above sea level.

That the pyroxenic rock once had the extension shown by the restoration in Fig. 68 is inferred from nearby laccoliths which have not been so greatly demolished by erosion. In some of these the pyroxenic rock still remains above the interior mass of syenite.

Associated complementary dikes. Another phenomenon, of the same category as those just described, is the common appearance of dikes in and around large intrusive bodies—batholiths and stocks of granite, syenite, diorite, etc.—where these bodies have become exposed by erosion. The dikes were injected after the main mass that they accompany had formed, and they cut this mass and also the surrounding rocks. In the surrounding rocks, the later intrusives may occur not only as dikes but also as sills and laccoliths. These rocks fall into two classes. In the first they are poor or entirely devoid of ferromagnesian minerals and are called *aplite* dikes, since the dikes of aplite that are generally associated with granites are the most common and best known representatives of this class. Aplites are also called *leucocratic* rocks (Greek *leukos*, white, and *kratein*, to rule) in allusion to their general light color resulting from the fact that they are composed mostly of feldspars or of feldspars and quartz. They are generally fine-grained rocks: some are of a sugar granular texture, some are aphanitic and to be classed as felsites. Some are porphyritic. They usually occur in narrow dikes, ranging from less than an inch in thickness up to a few feet.

The rocks of the second class are heavy, dark or even black, of basaltic aspect and composed chiefly of ferromagnesian minerals—biotite, hornblende, pyroxene, and olivine—in variable amounts and with subordinate feldspar. They are commonly porphyritic, having phenocrysts of the dark minerals in an aphanitic dark groundmass. Such rocks are called *lamprophyres* (Greek, *lampros*, shining, in allusion to the profusion of shining flakes of biotite in minette, the most typical of the lamprophyres); they are also termed *melanocratic* rocks, or rocks in which the dark minerals predominate (Greek *melanos*, black, and *kratein*, to rule). In our field classification they would be named biotite melaphyre (or mica trap), hornblende melaphyre, etc., according to the prevailing phenocrysts. They also occur mainly as narrow dikes.

These two kinds of rocks—the aplitic, light-colored and feldspathic, and the lamprophyric, dark-colored, and rich in ferromagnesian minerals—are termed *complementary* because taken together they represent the composition of the main masses they accompany. If we could mix them in amounts proportional to the bulk of their occurrence, we should obtain a rock whose chemical composition is that of these larger masses upon which they appear to depend as satellitic bodies. For some occurrences this has been actually tested and proved. When all the facts as to the mode of occurrence of these satellitic dikes are taken into account, they appear to have been formed by later intrusions from the same magmatic reservoir that yielded the magma of the larger masses, termed the "parent masses." In some way the parent magma has divided into two unlike submagmas. If these submagmas should break through to the surface, they would give rise to dissimilar lava flows as, for example, to felsites and basalts. They thus explain in part the phenomena noticed at many volcanoes, namely that lavas of several kinds have been erupted by one and the same volcano. The imposing extinct volcano of San Francisco Mountain on the plateau of northern Arizona erupted lavas of no less than five different kinds.

Not all dikes belong in the category of complementary rocks. On the contrary, the same magma that produced stocks and batholiths occurs also as dikes. These dikes have the same mineral composition as the larger masses. Only, as explained on page 123, they are likely to differ in texture from the rock of the stocks and batholiths, as a rule being porphyritic. This difference is the result of the different physical conditions (more rapid cooling and easier escape of the dissolved volatile matter) under which the magma injected into the dike fissures solidified, compared to the conditions under which the magma solidifies within the parent mass.

Magmatic differentiation. The varied lavas of volcanoes, the zoned laccoliths, the marginal zones of stocks and batholiths, and the associated complementary rocks described in foregoing sections present to us a body of geologic facts that can be satisfactorily explained only by the assumption that the magmas from which igneous rocks have formed had the capacity of developing derivative magmas that are unlike the original, or parent magmas. The process by which this is done is called magmatic differentiation. During differentiation there seem to be in general two opposite poles towards which the derivative magmas tend: one where the iron, magnesia, and to a large extent the lime are concentrated; the other where the alkalies, alumina, and the silica are concentrated. The one produces rocks rich in ferromagnesian minerals, such as gabbro; the other produces feldspathic rocks, such as granite. While this is the general description of the process, the details in nature are infinitely varied; thus in some places one can find a further, differential separation of the alkalies and an enrichment of potash in some rocks as compared with soda.

If a magma differentiates after it has come to rest in its chamber and while it is solidifying, the resulting rock body must evidently differ in its different parts. The most illuminating examples of differentiation in place are supplied by some of the world's thicker gabbroic sills. The Palisade sill along the Hudson River is a classic example. The main bulk of the sill is a coarse dolerite, but a short distance above the floor of the sill is a 50-foot layer rich in olivine. The explanation in brief is this. After the magma had been injected to form the sill, it began to cool. Olivine was the first mineral to separate from the magma, and being heavier than the magma the crystals sank and gradually built up a layer rich in olivine just above the bottom of the sill. The differentiation in the Palisade sill was manifestly effected by the crystallization of the olivine from the magma; hence the process is termed crystallization-differentiation.

The reality of magmatic differentiation as a process is well established on geologic grounds by a large body of facts. To realize that such a process has taken place and to understand how and why it has taken place are two entirely different matters. Although all who are thoroughly conversant with the facts are obliged to admit that it has taken place, a wide diversity of views prevails in regard to the manner in which the differentiation in general is effected, owing to the complexity of the phenomena and to insufficient knowledge. Few examples, unfortunately, are as clear as that of the Palisade sill.

Border facies by assimilation. Some border facies have been formed by magma spalling off fragments from its containing walls and incorporating this foreign material. The incorporated material may become mechanically disaggregated and scattered more or less homogeneously through the magma, or it may be acted on by the magma, or both processes may occur. The magma having thus been modified by assimilation, the resulting rock would differ in mineral composition from the interior, main mass. Some border zones have been formed this way; but this explanation does not apply everywhere, because many a border zone is found to have a mineral composition entirely different from that to be expected if it had assimilated the rocks with which it came in contact.

General explanation. The two hypotheses of magmatic differentiation just discussed do not give a general explanation for the differentiation of igneous rocks. The occurrence of complementary dikes, the eruption of lavas of several kinds from the same volcano, and the variety of different types that are not marginal in the same igneous mass, as well as other facts, show clearly that differentiation is a process that takes place in a magma so as to produce separate bodies of differing *liquids* that were independently injectible or eruptible. While we see that this is so, both from geological and chemical facts, no general explanation that is satisfactory in all respects has been offered for this process. Several hypotheses, which it would be beyond the limits of this work to state and discuss, have been suggested, but our knowledge is too limited as yet to enable us to estimate the proper value and applicability of these hypotheses.

Petrographic provinces. Consanguinity. The igneous rocks of certain regions, regardless of mode of occurrence and general composition, have particular features that are more or less distinctive of all members of the group that occur there. This community of characters is shown by the presence in all of the igneous rocks of particular varieties of minerals, or by peculiar textures, or by peculiarities of chemical composition and generally by a combination of these things. These characters are shown in varying degrees by all the different igneous rocks of the region: for example, by syenites which are chiefly composed of feldspar and by dolerites in which ferromagnesian minerals prevail; by intrusive stocks of granular rocks with their associated complementary dikes and sheets; and by lava flows of felsites and basalts. The common characters of the igneous rocks of certain regions are strongly marked, but in others they are to be seen only by the experienced observer. The fact that such distinguishing features occur in all the different igneous rocks of a certain region suggests that they are related to one another and were derived from a parent magma by differentiation. This relationship is termed the *consanguinity* of igneous rocks, and a region throughout which the rocks are thus genetically related is called a *petrographic province.*

Certain regions contain two or more consanguineous series of igneous rocks that were successively formed during epochs of igneous activity widely separated in geologic time. A period in which a genetically related series of igneous rocks was formed is called *petrogenic epoch,* and as this concept is more definite than "petrographic province," it is supplanting the older idea. Thus the Oslo region of southern Norway is characterized by the high percentage of soda contained in its magmas, which gave rise to certain minerals and peculiar rock textures; the magmas of Italy and central Montana were high in potash, which caused the formation of the mineral leucite, common in such regions but rare or unknown elsewhere; and the magmas of the western Mediterranean islands and eastern Spain were characterized by an abnormally high amount of titanic oxide.

Evidence of the consanguinity of a series of rocks cannot as a rule be seen during field work and by the megascopic study of rocks. It generally demands careful and complete investigation of thin sections under the microscope, aided by chemical analyses in the laboratory, together with a broad acquaintance with the literature of this subject, in order to be perceived and appreciated. The matter, however, is of great interest, and although one may not be either a chemist or petrographer, he may nevertheless appreciate the significance of its bearing on problems of the greatest importance in geology. Manifestly, before we can safely theorize as to the origin and history of the Earth we must first know the nature of its component parts and the laws governing their distribution.

EPIMAGMATIC PROCESSES

When a body of magma has become emplaced in the position that it is destined thenceforth to occupy as a solidified rock mass, it begins to cool and eventually minerals crystallize out. During the period of crystallization the volatile substances dissolved in the magma and hence held in it under pressure, such as fluorine, boric acid, carbon dioxide, and especially and chiefly water, are gradually excluded, except in so far as they become part of the chemical composition of some of the minerals.

This period in the life of a magmatic body, when it is solidifying and giving off heat and hot gases as its minerals crystallize, is called the pneumatolytic period (from the Greek *pneuma,* gas, and *lysis*, a loosening), and these agents generate important results. At the surface the emanating gases give rise to fumaroles, solfataras, and hot springs;

in the depths they aid the heat that is given off in producing in the rocks that surround the igneous mass the transformations included under the term of *contact metamorphism*; and in the already solidified parts of the igneous mass they bring about the formation of pegmatites and in some places of ore deposits. These processes are termed *epimagmatic* because they take place at or near the end of the life of the magmatic body.

Pegmatites. It has previously been stated that stocks and batholiths of granite, syenite, diorite, etc., that have been exposed by erosion are cut by complementary dikes of aplite and lamprophyre. Many of them are also cut by dikelike masses with very striking characters, the most marked of which is the large size of their constituent minerals. Such masses are called *pegmatite* dikes or veins (it is not always possible to determine whether a given pegmatite is a dike or a vein), from the name originally given to the remarkable "graphic" intergrowths of quartz and feldspar in many of the pegmatites associated with granites. Pegmatites cut not only the parental masses, the stocks and batholiths to which they owe their origin, but also the enveloping rock masses. The features that particularly characterize pegmatites are as follows:

A. They consist mainly of the ordinary minerals that compose the granite or similar rock with which they are genetically associated. These minerals, instead of having the regular order of successive crystallization, show by their interpenetration that they have crystallized nearly, if not exactly, simultaneously.* The large size of the individual crystals is a striking character that has already been mentioned. Feldspar and quartz may occur in crystals a foot or even several feet long, apatite in dimensions like the handle of a broom, muscovite in crystals yielding plates from a foot to ten feet in diameter, and other minerals in similarly impressive sizes. These sizes are not the averages; they are the extremes, which, however, are not infrequently attained.

B. Another peculiar feature is that the minerals in many pegmatite masses tended to grow outward from both walls of the mass and to project inward toward the center. There may even be an empty space at the center of the pegmatite mass into which the minerals project, showing crystal faces as in miarolitic cavities (page 131) or in the vugs of minerals veins. The whole effect is to produce a crude banding or ribbon structure,† like that so commonly seen in mineral veins.

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* See in this connection the description of graphic granite in the granite pegmatite veins, page 174.

† The most recent designation for this long-known feature is "zoning."

Pegmatites of this kind are without doubt veins that were formed at high temperatures—from about 600° C down to 400° C: they are hypothermal veins.

C. Another character is the extreme variability from place to place in the relative proportions of the component minerals, a variability not seen in the main igneous rock mass with which the pegmatite is associated. Thus masses of granite pegmatite when traced along their strikes may vary greatly in the relative porportions of quartz and feldspar. Traced outward from the parent mass into the inclosing rocks, they may even grade into pure quartz veins. Other pegmatites grade into fine-grained granite (aplite) or felsite, and this gradation may take place very abruptly.

D. They are likely to contain minerals that either do not occur at all in the parent rock mass or, as shown by microscopic examination, are only sparingly distributed in the parent mass in minute crystals. These minerals fall into two classes. In the one they are made up of the ordinary oxides that compose magmas—alumina, lime, iron, soda, etc.—plus the volatile elements that were dissolved in the magma. Abundant are tourmaline, which shows that boron was present; topaz and fluorite, which demonstrate that fluorine was present; and many minerals, of which muscovite is perhaps the most prominent, that prove the important role played by water. The minerals of the other class are characterized by rare elements, such as lithium, cesium, beryllium, uranium, molybdenum, cerium, columbium, tantalum, and others—elements that occur only in minute traces or not at all in the parent rock mass. In combination with silica, titanic oxide, phosphoric acid, zirconia, carbonic acid, and fluorine, these elements give rise to a host of minerals too numerous to mention, but beryl and spodumene can be cited as examples. No sharp distinction can be drawn between these two classes of minerals; many can be placed in either class, but definite types of both, like those mentioned, can readily be found.

These minerals generally occur in crystals of great size, even of gigantic size, and in some pegmatites they are aggregated in large amount. Because of the great variety of minerals and the large size and perfection of the crystals, pegmatite dikes and veins are mineralogically of great interest and are therefore favorite hunting grounds of mineral collectors.

Each kind of magma (or rock) is characterized by its own special mineral assemblages in its pegmatite dikes. This rule applies not only to the ordinary minerals that compose the pegmatites and distinguish them from one another but to the accessory minerals as well. Thus the minerals of pegmatites that are genetically related to granites differ greatly from the minerals that occur in the pegmatites genetically related to nepheline syenites, as these in turn differ from those associated with gabbros. Therefore there are many kinds of pegmatites, but when "pegmatite" is used without qualification it means granite pegmatite, because pegmatites are far more abundantly associated with granites than with other phanerites.

Origin of pegmatites. The simultaneous method of crystallization, the rudely layered arrangement of the minerals along the walls of some pegmatites, the inconstant distribution of the component minerals, and their frequent huge size all show that the conditions that prevailed while a pegmatite of this kind formed differed from those which attended the solidification of the main rock masses. The hydroxyl, fluorine, and boron content of the pegmatitic minerals shows that volatile constituents were present. Bearing these facts and those of the geologic mode of occurrence in mind, we can present to ourselves a view of the origin of pegmatites somewhat as follows:

When a body of magma, such as will form a stock or batholith, has become emplaced in the crust it commences to cool. This cooling takes place of course first in the upper and outer portions, and here the mass begins to solidify by crystallization. As it becomes solid, it breaks up by contraction into jointed masses. As the cooling and the crystallization of minerals from the magma go on, the volatile components of the magma, beyond those retained by the minerals that use them in their chemical composition, become concentrated in the still liquid portion of the magma, and as they accumulate more and more their pressure increases enormously Consequently they tend to force their way upward along the fissures in the solidified parts above and at the sides. These fissures thus become channel ways for the residual magma, whether differentiated or not, and they afford avenues of escape also for the gases that collect in them and in the cracks in the surrounding shell or hood of already formed rocks, under pressures that must generally be enormous.

At 374° C water reaches its critical temperature; that is, at this temperature or above, its vapor cannot be turned back into liquid merely by pressure, however great the pressure may be. At the critical temperature its expansive force exceeds 3000 pounds to the square inch—a pressure equal to the weight of a column of granite rock 2500 feet high. As the temperatures of magmas greatly exceed the critical temperature of water, water must therefore be in the gaseous state when it escapes from the magma in which it was contained. Such superheated steam, although its temperature is above the critical temperature, can dissolve and carry silica and other substances, as has been experimentally established.

From a magma highly charged with water the minerals would crystallize somewhat differently from the way they would from normal magma. In some places the magma might have only moderate water content and the results would approximate those obtaining in the main body of magma. If the water content diminished by escaping into the wall rocks, the magma would solidify as an aplite or felsite. With increased content in another place, the magma might pass from the state of a liquid into a gaseous or aqueous solution and here would be favorable conditions for the growth of crystals to gigantic size and for their growth perpendicular to the walls of dike-like masses. Finally, passing onward, the solution may become depleted of most of its constituents, only silica would be carried, and the resulting deposit might grade into a quartz vein. Thus, we can visualize how dikes of igneous rock, pegmatites of variable composition, and quartz veins may be formed.

Many granite pegmatites show the results of extensive replacement. Most commonly the microcline has been replaced by snow-white albite. The albitization may be so thorough that little or no microcline is left. The albitization may be succeeded by replacement by lithium minerals, such as lepidolite or spodumene. Although it has been estimated that in general not more than one per cent of pegmatites have been affected by replacement, it is that one per cent that carries all the remarkable minerals for which pegmatites are famous.

Contact metamorphism. The term *contact metamorphism* is applied to the changes caused by a body of magma that has come in contact with other rocks already formed. The word metamorphism means a change of form or body and is applied to those results of geologic factors acting within the Earth's crust, by which rocks are reorganized by the formation of new minerals and textures. Only the metamorphism caused by magma is described here; metamorphism of other kinds, by which rocks are changed over wide areas through various geologic agents is considered in a later chapter. The results of the two kinds of metamorphism are alike in many respects, but contact metamorphism is far simpler to understand.

The contact-metamorphic effects produced by a body of magma are manifested in two ways: one, *endogene* effects: internal changes, by which the normal character of the igneous rock itself has been changed; two, *exogene* effects: external changes, produced in the rocks in contact with the igneous mass. We shall consider the internal or endogene effects first.

Endogene Effects. It has already been pointed out that the border zone of an intrusive mass may differ in mineral composition from that of the mass as a whole. In places, even where such border facies have not developed, more or less of a change in the minerals of the igneous rock can be seen at the immediate contact or near it. These changes were produced by the action of the gases that were liberated, as previously mentioned, as the igneous body cooled and crystallized. The liberated gases escaped to the margin of the igneous body and into surrounding rocks. The chemical activity of these hot gases formed minerals that are not generally present in the main part of the igneous body but are characteristic of pegmatites. In granites the most characteristic is perhaps tourmaline, which indicates that boron, hydroxyl, and fluorine were present in the metamorphosing solutions. Tourmaline tends to occur in the place of the biotite in the main igneous body, and its occurrence as a component of the granite should always lead one to infer nearness to the contact, though it is also found near fissures that served as the conduits for pneumatolytic exhalations.

The granite of Harney Peak in the Black Hills illustrates this feature; besides the usual quartz and feldspar, the granite contains black tourmaline, abundant muscovite, green beryl, and red garnets: these minerals recall the associations seen in pegmatites.

The accumulation of dissolved gases may have become so great at the border of an igneous mass before crystallization set in that the conditions were favorable there to form a zone of pegmatite. Granite stocks have thus become wrapped by a mantle of pegmatite; the large plates of muscovite are set perpendicular to the contact and the mixture is much enriched in quartz. Pegmatite mantles of the kind just described are especially developed around the intrusive masses such as stocks that are cupolas on the backs of batholiths. The small granite stocks of Saxony and Bohemia, whose magmas were heavily charged with gases, are the classic examples.

A much more common endogene effect is a change in texture. This effect is independent of any change in mineral composition; in fact, it occurs chiefly where the mineral composition remains constant. The most common effect of this kind in a rock is a change in the average grain size, which grows smaller toward the contact. The rock may indeed be exceedingly fine grained at the contact; for instance, a granite averaging three millimeters in grain size may at the contact be an aphanitic felsite. Such a fine-grained border facies is the result of simultaneous crystallization around a large number of nuclei, produced by the chilling effect of the contact wall. Examples are even known in which the chilling caused by the cold rocks with which the magma came in contact was so drastic that the border zone solidified before any crystallization could begin, and consequently a thin selvage of glass was produced. Such development of glass is most common in narrow dikes, in which the cooling effect of the contiguous rock is most strongly felt.

Moreover, a fine-grained border facies may contain large distinct crystals, or phenocrysts, and thus be porphyritic, although the main mass into which it grades is of equigranular fabric. The phenocrysts may have been formed before the magma came to rest; in the main mass they may be of the same size as the rest of the later rock grains, but at the contact their contrast with the finer-grained material in which they are embedded produces a porphyry. On the other hand, in many porphyritic dikes and sills phenocrysts are entirely absent from the border zones. In a dike of this kind the phenocrysts must therefore have formed after the magma had been injected into the dike fissure, because if they had been brought up in the ascending magma they would occur at the contacts as well as in the interior of the dike.

Exogene effects of contact metamorphism in general. The effect of the heat and gases given off by an intrusive mass of magma on the surrounding rocks with which it is in contact varies with a number of factors. For one thing, the effect varies with the size of the intrusive mass. A practical application of this rule is this. In a coal field cut by dikes, the amount of coal commercially ruined by a dike cutting through a coal bed is roughly equal to the thickness of the dike; accordingly a 10-foot dike spoils twice as much coal as a 5-foot dike. To stocks and batholiths the rule cannot be so simply applied. The exogene effect varies also with the nature of the gases that are given off, as described under the origin of pegmatites. Another factor is the nature of the rock that is being affected, some kinds being more easily metamorphosed than others. The extent of the metamorphism depends also on the attitude of these rocks; that is, in sedimentary beds, it depends on the position of the bedding planes with reference to the igneous mass. All these are important factors and each deserves separate treatment in order that the subject may be fully understood.

In general, the most noticeable field evidence of external contact metamorphism ("exomorphism") is a baking or hardening of the surrounding rocks. The result is that the rocks thus hardened generally resist erosion better than the intrusive mass or the unchanged enveloping rocks and consequently give rise to distinct projecting topographic forms. This is admirably illustrated in the Crazy Mountains of Montana, where the resistant rocks of the contact zone project as high ridges and peaks that encircle a more deeply eroded mass of intrusive igneous rock and rise sharply from a sloping plain of soft unchanged shales and sandstones. In similar fashion a dike and the surrounding beds may be lowered more rapily by erosion, while the contact-altered walls on both sides of the dike are left projecting as two parallel ridges.

The mineralogical effect of external contact metamorphism is that, in general, where the agents have made themselves most strongly felt the rocks have been thoroughly transformed. This transformation is produced by an interchange of the ions of adjacent minerals whereby the old minerals are broken up and new minerals are formed. The bulk chemical composition of the altered rock generally remains the same as before metamorphism, except that volatile compounds—water, carbon dioxide, and organic matter—are driven out, and in some cases, volatile components—fluorine, boron, etc.—may be added by the gases escaping from the igneous mass.

Modes of occurrence. The widest and most pronounced contact-metamorphic zones occur around stocks and batholiths. This is most natural, because large igneous masses can supply immense amounts of heat and gases. Around many stocks and batholiths, contact-metamorphic zones as much as a mile or more in width have been formed. After those found in such situations, perhaps the most striking are those around old volcanic necks. The breadth and intensity here generally seem disproportionate to the size of the igneous mass, but this large amount of metamorphism is explained by the fact that volcanic necks represent conduits through which molten matter had long been passing. The continuous supply of magma in the conduit and therefore of heat has had a cumulative effect. In such places there is likely to be no endogene effect of cooling on texture as described above: the conduit walls became so highly heated that the texture of the igneous rock remains the same up to the very contact wall.

Dikes show considerable diversity in their metamorphic powers. Small dikes have exerted an effect for only a few inches or even less, whereas large ones may have caused changes for many yards on either side. But the reverse is also true. Again, some dikes have served as conduits for magma passing up through them into large intrusive masses higher in the crust, supplying magma to sills or laccoliths or issuing at the surface as extrusive outflows. Around them the metamorphism is naturally greater, other things being equal, than where a fissure was filled by a single charge of magma that immediately came to rest and solidified. For this reason the metamorphism induced by sills and laccoliths is generally inconsiderable, since they represent but a single charge of magma into the rocks surrounding them, which is not renewed. Immediately at the contact and for a few feet or yards beyond the rocks may be altered, but the effect dies out abruptly, except near very thick sills and large laccoliths. Lava flows cause at most a small amount of baking or hardening of the rocks and soils on which they rest, but in general the effects are so small as not to be noticeable.

Effect of position of the rocks. The width of the contact-metamorphic zone varies considerably from place to place about the intrusive mass. This variation may be due to underground irregularities in the igneous rock body: a broadening of the zone, for example, indicates a corresponding extension of the igneous mass below, as illustrated in Figs. 69 and 70.

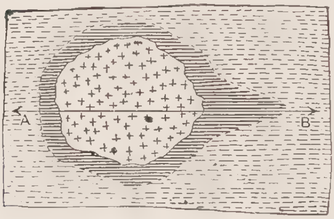

Fig. 69. Ground plan or map of a stock and its contact-metamorphic zone.

In bedded rocks the position or attitude of the planes of bedding relative to the intrusive mass is also important. Thus in Fig. 71 the beds at *B* dipping into the mass of granite *C* tend to have their bedding planes and joints opened by the upward movement of the magma, and their position is such as to facilitate the entrance and migration of the gases and heat, thus producing a broad contact zone. On the side *A*, on the contrary, the conditions are just the reverse, and a much narrower contact zone is the result.

Fig. 70. Vertical section along line *AB* in Fig. 69.

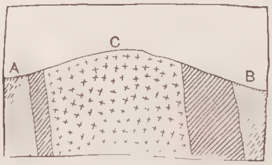

Fig. 71. Section showing that the width of the contact-metamorphic zone is influenced by the dip of the beds with reference to the contact surface.

Effect on rocks of different kinds. The most notable effects of contact metamorphism are produced on sedimentary rocks—sandstones, dolomites, limestones, and shales or slates. On pure *quartz sandstones* the effect is relatively slight, though in the innermost zone adjacent to the igneous mass they are likely to be converted for a short distance into quartzites. Pure *limestones* are recrystallized and changed into *marble*, not infrequently in large masses and through considerable distances from the igneous contact. The most striking effects are produced in impure limestones containing quartz sand and clay mixed with them. In such limestones the quartz drives out the carbon dioxide, and the carbonates are changed to silicates. If the limestone is dolomitic, and hence contains magnesia, the results are more complex.

Some of the simpler of the metamorphic changes are readily shown by equations that represent the chemical changes involved.

$$\overset{\text{Calcite}}{CaCO_3} + \overset{\text{Quartz}}{SiO_2} = \overset{\text{Wollastonite}}{CaSiO_3} + \overset{\text{Carbon dioxide}}{CO_2}$$

$$\overset{\text{Dolomite}}{CaMg(CO_3)_2} + \overset{\text{Quartz}}{2SiO_2} = \overset{\text{Pyroxene}}{CaMg(SiO_3)_2} + \overset{\text{Carbon dioxide}}{2CO_2}$$

$$\overset{\text{Calcite}}{3CaCO_3} + \overset{\text{Kaolinite}}{H_4Al_2Si_2O_9} + \overset{\text{Quartz}}{SiO_2} = \overset{\text{Garnet}}{Ca_3Al_2(SiO_4)_3} + \overset{\text{Carbon dioxide}}{3CO_2} + \overset{\text{Water}}{2H_2O}$$

$$\overset{\text{Calcite}}{CaCO_3} + \overset{\text{Kaolinite}}{H_4Al_2Si_2O_9} = \overset{\text{Anorthite}}{CaAl_2Si_2O_8} + CO_2 + 2H_2O$$

In some contact zones the original rock was thus entirely changed into silicates or mixtures of them, but generally the metamorphic rock is an impure marble or altered limestone containing the newly formed minerals aggregated into lumps or bunches. If carbonaceous material was present, it has commonly been driven off; rarely it stays in the rock and is changed into graphite.

Besides the minerals mentioned, many others, the origin of which is due to substances brought by gases given off by the igneous rock, may also be formed, such as mica (phlogopite), chondrodite, hornblende, vesuvianite, epidote, and tourmaline. The main components of these minerals were already present in the original rock; but the gases have supplied certain components—hydroxyl, boron, and fluorine—needed to form the new minerals. Such minerals furnish transitions to the more typical pneumatolytic contacts mentioned below. Many of these minerals contain iron (ferrous, ferric, or both), and this iron came from the limonite or other hydrated iron oxides mixed in with the impure marly beds and deposited with the other material at the time of their formation.

In some contact-metamorphic zones, however, iron has been added by hot gases expelled from the magma; it was carried there probably in the form of a volatile chloride or fluoride. Perhaps, of the minerals mentioned, garnets, pyroxenes, vesuvianite, and epidote are the most customary in altered limestones of contact-metamorphic zones. Such zones are known in many parts of the world, and some of them have become famous for the variety and beauty of their crystallized minerals.

Shales and *slates* generally show well-marked and characteristic effects of contact metamorphism. Although shales and slates consist mostly of microscopic fragments of quartz, flakes of kaolinite, mica, and chlorite, they have a considerable range in composition and, accordingly, a diversity in the results of metamorphism. Some have been baked into a hard aphanitic rock called *hornfels*; it has conchoidal fracture, is black or dark gray, and closely resembles fine-grained black basalt.

If the shales and slates were rich in kaolinite, andalusite commonly has formed, probably according to this formula:

$$\overset{\text{Kaolinite}}{H_4Al_2Si_2O_9} = \overset{\text{Andalusite}}{Al_2SiO_5} + \overset{\text{[Quartz}}{SiO_2} + \overset{\text{Water}}{2H_2O}$$

At the contact a rock composed of andalusite, commonly in recognizable grains and prisms, quartz, and biotite in glimmering specks forms a finely granular rock, as a rule dark gray or black and much resembling some andesites. If the original rock was a slate, the cleavage has now been completely obliterated. The newly formed rock is an *andalusite hornfels*. Farther from the contact the hornfels loses its granular texture; it becomes more schistlike or perhaps slaty but is dotted with andalusite prisms. These prisms generally contain within themselves carbonaceous matter arranged as crosses in the remarkable manner characteristic of andalusite, forming the variety of andalusite known as chiastolite, as mentioned on p. 59.

Still farther from the contact effects of metamorphism are less and less marked, and the rocks retain more and more of their original characters. This fading out of the metamorphic effects away from the contact has caused the contact-metamorphic zone to be called the contact-metamorphic *aureole*. In the outermost part of the aureole the most evident effect is a spotting of the shales or slates as a result of the incipient development of cordierite. Such knots are characteristic of mild contact metamorphism and, when met in the field, should always lead to search for more intense effects and the possible nearness of an intrusive igneous rock mass. The intrusive mass may, of course, be below and not yet exposed by erosion.

Just as all kinds of gradations between sandstones, limestones, and shales occur in nature, so the many varieties of rocks produced by contact metamorphism, as described above, vary and grade into one another.

In the nature of things, already existent igneous rocks are less susceptible to contact metamorphism than are the sedimentites. Granites and other feldspathic rocks are nearly immune to change by contact metamorphism. Ferromagnesian rocks, however, especially those composed chiefly of pyroxene, show in places considerable effects. The pyroxene is converted into hornblende and the rock becomes an amphibolite and even in places a glistening black hornblende schist.

Pneumatolytic contacts. It was mentioned above, in connection with the changes observed in limestones, that some of the newly formed minerals were formed by reactions with the gases given off from the igneous mass. In some contact zones the rocks have been converted into masses of such minerals, eloquent testimony to the abundance and energetic action of the gases given off by the solidifying magma. Such minerals as tourmaline, topaz, fluorite, vesuvianite, and mica (muscovite), which contain hydroxyl, fluorine, and boron, are characteristic of these occurrences. The masses of metamorphic minerals thus formed do not make a continuous border around the igneous body but appear here and there, especially along fissures. The controlling factors are the avenues of escape of the gases from the solidifying magma and the reactivity of the rocks through which the gases are streaming.

Contact-metamorphic zones and ore bodies. The contact-metamorphic zone that surrounds an igneous mass is an especially favorable place for the deposition of certain kinds of ores. Metal-laden gases escaping from the magma travel outward through the zone, and, by reaction with the rocks and by losing temperature as they move outward, they deposit their metallic content. As a result many valuable ore deposits of iron, copper, tungsten, and other substances have been formed by the action of magmatic gases in such contact zones. They are called *contact-metamorphic ore deposits.* They are generally restricted to limestones, because limestones react easily with the magmatic gases and cause the gases to deposit the iron, copper, and other substances they were carrying. To treat this subject adequately would carry us far beyond the limits of this work, but further information can be found in the treatises that deal with the origin of ore deposits.

CHAPTER VII

CLASSIFICATION OF IGNEOUS ROCKS

Introductory. The igneous rock masses of the Earth have certain features that can be used to distinguish them, one from another, such as their geologic mode of occurrence, their mineral composition, their texture, and their chemical composition. Brief inspection serves to show, however, that all these features are gradational, so that no hard and fast lines of separation can be found. If we classify igneous rocks according to their geologic mode of occurrence and therefore divide them into intrusive and extrusive rocks, then obviously every lava flow is (or was once) prolonged into depths below by an intrusive continuation in the form of a dike or volcanic neck. We should have to separate the intrusive from the extrusive at some point by an arbitrary plane; the rock above this plane would receive one name, the rock below it another, though manifestly the material just above the plane and that just below are absolutely alike. Similar difficulties are met when we consider the other features mentioned: rocks grade into each other mineralogically, chemically, and texturally, and if lines are drawn, they must be drawn arbitrarily. Owing to these facts, much diversity of opinion exists as to how igneous rocks should be classified, some authorities laying stress on one feature, some on another. By general consent among petrographers, especially since the use of the microscope has revealed the composition of the aphanitic rocks, many kinds of igneous rocks have been recognized, differentiated primarily by their textures and by the kinds and relative quantities of their component minerals.

Older megascopic classification. Before the microscope came into use in studying rocks, they were divided into two groups: those whose component mineral grains are large enough to be seen and recognized—phanerocrystalline rocks; and those too fine-grained to permit recognition—aphanitic rocks. The phanerocrystalline group was subdivided on the basis of the minerals present, and the aphanitic group was subdivided according to color, texture, luster, and other physical properties. In this way a megascopic classification came about, extremely useful for geologic and common purposes, which gave rise to such terms as granite, diorite, porphyry, greenstone, basalt, and others.

Effect of the microscope. When the petrographic microscope came into use, it was discovered that the aphanitic rocks could be studied, and that their component mineral grains could be determined almost as easily as those of the coarse-grained ones. The results of these studies showed that rocks are vastly more diverse than had been suspected. The microscope revealed also that the coarser-grained rocks contain many minerals not previously known to occur in them. It showed also great variations among the minerals that compose them and disclosed many differences in texture. To express these differences among the rocks and to connote the ideas that these differences engendered, not only has a host of new rock names been invented, but also the old megascopic terms have been defined and redefined by various authorities, until nearly all have lost their original significance.

This has been an unfortunate phase of the history of the development of petrography as a science, because the former megascopic or field names, as we should now term them, served a useful and necessary purpose, which the more exact and scientific nomenclature of modern petrography cannot serve. The person who desires to deal with rocks and name them megascopically, such as the field geologist, the engineer, and the architect, is left without the equipment for doing so. A single illustration will suffice. The old term granite meant any granular igneous rock, and then later one composed of quartz and feldspar. Now, according to modern usage, a rock is not a granite unless the feldspar is chiefly potassium feldspar, but if its feldspar is dominantly sodic plagioclase, it is termed a granodiorite. This distinction based on the nature of the feldspars is one that ordinarily cannot be made without microscopic study.

The redefinition and specializing of the useful general field terms has very much the same result as if botanists had redefined such as bush, tree, vine, and shrub and had made them the names of particular species or genera, so that if tree, for instance, were properly used, it would designate only oaks, or even *Quercus alba.*

Present need for a field classification. It is clear that the parting of the ways has long been reached, and it ought to be definitely recognized that the further scientific development of petrology and of the classification and nomenclature of rocks must be left to petrographers, and that those who have occasion to deal with rocks in the purely megascopic manner must have a method of classification and a set of terms of a somewhat different scope and usage. They must in large measure revert to that which was in vogue before the microscope came into use.

It matters little whether or not such a classification is completely consistent with all the principles that the study of rocks has revealed. To be useful the classification must be practical, and to be practical it must be based entirely on the evident, megascopic characters of rocks, such as can be seen by the unaided eye or pocket lens.

Classification used in this book. As the object of this book is to treat rocks megascopically, the following classification is used. Igneous rocks are subdivided first on the basis of their grain size into four major classes. In order of decreasing grain size these classes are *phanerites*, *microphanerites*, *aphanites*, and *glasses*.

I. *Phanerites.* Phanerites are those rocks in which more than 50 per cent of component minerals are large and distinct enough to be seen and recognized by the unaided eye or with the lens. Such rocks are said to be *phanerocrystalline* (Greek *phaneros*, visible, manifest). Phanerites are predominantly equigranular rocks whose average grain size exceeds 1 millimeter in diameter.

II. *Microphanerites.* This class conveniently includes those fine-grained equigranular rocks in the twilight zone between the phanerites and aphanites.* As an important subgroup II*B*, microphanerites carrying phenocrysts are recognized as *microphanerite porphyries.*

III. *Aphanites.* Aphanites (meaning, "not phanerites") are those rocks that have less than 50 per cent of phanerocrystalline components. The normal aphanite is more or less porphyritic, having phenocrysts scattered through an aphanitic† groundmass. The groundmass is of dull stony appearance; it is homogeneous and seems to be composed of a single substance, although made up as a rule of a minutely crystalline aggregate of more than one mineral. Since most aphanites are porphyritic, a more significant name for this group is *aphanite porphyries*, which can be shortened to *aphanophyres.*

IV. *Glasses.* Glasses are rocks wholly or largely made up of glass. They are recognizable by their luster, which ranges from vitreous to pitchy; by their conchoidal fracture; and by other characters. They can be considered to be rocks of zero grain size.

Subdivisions of the phanerites. All of the igneous rocks having been divided into four major classes on the basis of grain size, it now remains to show how these major classes can be subdivided. The

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* This usage was advocated as long ago as 1910. (Joseph Barrell, The lithology of Connecticut. *Connecticut Geol. and Nat. Hist. Survey, Bull.* 13, 1910.)

† "Dense" is often used incorrectly in geology as a synonym for "aphanitic." Dense is used also in geology to mean (1) closely set, crowded; (2) compact, i.e., devoid of vesicles or pore spaces; or (3) having a high density (high weight per unit volume).

phanerites, as already suggested, are subdivided on the basis of their mineral composition. First, they are divided into two main groups:

A. Phanerites in which feldspars or feldspars and quartz predominate: feldspathic rocks.

B. Phanerites in which ferromagnesian minerals (olivine, pyroxene, hornblende, biotite) predominate: ferromagnesian rocks.

The feldspathic rocks as a rule are light-colored: white, gray, or red, but this rule is not absolute, as some feldspars are very dark from included pigment. The ferromagnesian rocks are generally dark or black, but this also is not invariably true, as some, such as those composed wholly of olivine, are rather light colored.

The feldspathic rocks (group *A*) are further divided on the basis of the proportion of quartz to the feldspars and on the kind of feldspar present, whether orthoclase or plagioclase. Those containing orthoclase and considerable quartz are called *granite* if they are equigranular, and *granite porphyry* if porphyritic; those in which quartz is absent or in minor quantity are called *syenite* and *syenite porphyry* respectively. The feldspathic rocks are subdivided into varieties on the basis of particular mineral characters as shown in their descriptions in the next chapter.

The ferromagnesian rocks (group *B*) are subdivided on the basis of the proportion of the feldspars to the ferromagnesian minerals into those in which feldspar is merely subordinate in amount to the ferromagnesian minerals, and those in which feldspar is absent. Thus we have:

1. Rocks in which ferromagnesian minerals predominate; feldspar subordinate.
2. Rocks consisting wholly of ferromagnesian minerals.

Phanerites composed chiefly of ferromagnesian minerals are termed gabbro; the feldspar in them is generally a calcic plagioclase.

Porphyries occur in the group of ferromagnesian rocks, but they are far less common than those in the feldspathic rocks. Rocks are known in which ferromagnesian minerals, other than hornblende and pyroxene, predominate over feldspar, but they are of minor importance in a megascopic classification and are therefore omitted.

The subdivision, phanerites, that consists wholly of ferromagnesian minerals, is further divided on the basis of which ferromagnesian mineral predominates. The rocks in which olivine predominates are termed *peridotites*. Those in which pyroxene predominates are called *pyroxenites;* and if hornblende predominates, they are called *hornblendites. Porphyritic* equivalents of any of these rarely occur.

In summation, in considering the classification of the phanerites, one should consult what was said about the chemical composition of igneous magmas and the variations in mineral composition, beginning on page 113. Phanerites, however, cannot be classified entirely for megascopic purposes by the diagram given on page 116, for, in general, we cannot discriminate by the unaided eye between the different kinds of feldspars. Thus the rock there shown as quartz diorite must be classified as granite, while, as compared with the diagram, the diorite and gabbro mentioned above broadly overlap. Still, in a general way, despite these exceptions, the classification, by distinguishing between the feldspathic and the ferromagnesian phanerites, brings out the ideas there expressed. Silicic phanerites are nearly all intrusive, but many subsilicic ("basic") phanerites occur both as intrusive and as extrusive masses.

MICROPHANERITES. Microphanerites, comprising class II of the igneous rocks, are intermediate in grain size between the phanerites and the aphanites. The rocks of this class are obviously crystalline to the unaided eye, but the constituent minerals are too small to be identified megascopically. They require to be examined under the microscope to be certainly identified. Hence the rocks are said to be microcrystalline and are called microphanerites.

SUBDIVISIONS OF THE APHANITES. In considering the third class of rocks, the aphanites, the grain size of which is so minute that the mineral grains cannot be perceived megascopically, we have little on which to base a field determination except color. The aphanites are accordingly divided into two groups: light colored and dark colored. If a rock is either white or black, of course no difficulty arises in assigning it to one or the other of the two groups, but all gradations of color exist and it would accordingly become a matter of pure choice to which group a particular rock would be assigned. Some closer definition of the terms is evidently needed. We may supply this as follows: *dark* includes rocks that are *dark gray*, *dark green*, or *black*; *light* includes all other colors—*white*, *red*, *purple*, *yellow*, *brown*, *light* and *medium gray*, and *light* and *medium green*. Light-colored aphanites are *felsites*, and the dark aphanites are *basalts*. The division thus made also expresses in a general way an important fact about their composition. The felsites are derived from magmas which, had the physical conditions been such as to produce coarser-grained rocks, would have formed granites and syenites. On the other hand, basalts are the aphanitic equivalents of the gabbros. Although exceptions are many, the rule holds true that the light-colored aphanites as defined above are chiefly feldspathic, the dark are mainly ferromagnesian.

Although some aphanites are of homogeneous texture and aspect, *i.e.*, completely aphanitic and devoid of phenocrysts, the great majority are porphyritic. We then have *felsite porphyry* (in shortened form, *felsophyre*) and *basalt porphyry* respectively, according to the color of the groundmass.

The aphanitic porphyries can be called in condensed form *aphanophyres*, and they can be subdivided respectively into *leucophyres* (light-colored *porphyries*) and *melaphyres* (dark-colored porphyries). These porphyries can be further subdivided and named according to the minerals that occur in them as prominent phenocrysts. Thus we can have *quartz felsite porphyry*, or *quartz felsophyre* (the rock so named is implied to be light colored, minutely grained, and containing phenocrysts of quartz), *feldspar felsophyre*, and *hornblende felsophyre*, or *quartz leucophyre*, *feldspar leucophyre*, and *hornblende leucophyre*. Similarly, the dark-colored porphyritic rocks comprise *augite melaphyre*, *biotite melaphyre*, *feldspar melaphyre*, and others. Descriptive names of this kind can easily be made, but those given should suffice as examples. The great advantage of this scheme of naming rocks is that it is elastic and descriptive names can be coined *ad libitum*. An even greater advantage is that it advertises the fact that it is a field classification.

However, it is often possible to subdivide the felsite porphyries according to the kinds of rocks that are commonly recognized in current geologic literature. The felsophyre group is so broad as to include, as shown in Table 1 (pages 162–163), rhyolite, trachyte, latite, dacite, and andesite. These various rocks can be distinguished megascopically with sufficient accuracy on the basis of the kind of phenocrysts present. If a felsophyre contains phenocrysts of quartz and feldspar, it is either a rhyolite or a dacite. If most of the feldspar phenocrysts are unstriated, thus indicating that orthoclase is the dominant feldspar, it is a rhyolite. If most of them are striated, indicating plagioclase, the rock is probably a dacite. If all the sialic phenocrysts are orthoclase, the rock is a trachyte. If many plagioclase phenocrysts occur with the orthoclase phenocrysts, the rock is a latite. If all the feldspar phenocrysts are plagioclase, it is an andesite. Biotite and hornblende, singly or together, may occur as phenocrysts along with the quartz and orthoclase in rhyolite. Biotite, hornblende, and pyroxene, singly or together, commonly occur with the sialic phenocrysts in dacites and andesites. Thus we can distinguish varieties of dacite and andesite: biotite dacite, hornblende dacite, pyroxene dacite; hornblende andesite, hornblende-pyroxene andesite, etc.

The distinction between rhyolite and dacite, and between trachyte, latite, and andesite cannot invariably be made with the unaided eye, since it depends on the kind of feldspar present, and in some rocks

TABLE 1 Classification

Major Classes	Subdivisions of Major			
	Light-Colored Rocks			
	Orthoclase,[1] Chief Feldspar; Sodic Plagioclase, Subordinate		Orthoclase [1] = or < Sodic Plagioclase	
	B H P		B H P	
	with quartz	without quartz	with quartz	without quartz
I. Phanerites (equigranular; grain size > 1 mm)	Granite	Syenite Nepheline syenite	Quartz Monzonite and Granodiorite	Monzonite
II*A*. Microphanerites (equigranular; grain size < 1 mm)	Microgranite Aplite	Microsyenite	Microquartz-monzonite and Microgranodiorite	Micromonzonite
II*B*. Microphanerite Porphyries	Granite Porphyry	Syenite Porphyry Nepheline syenite porphyry	Quartz Monzonite Porphyry and Granodiorite porphyry	Monzonite Porphyry
III. Aphanites and Aphanite Porphyries	Rhyolite	Trachyte Phonolite	Quartz latite	Latite
	Felsite and Felsophyre			
IV. Glasses (composed wholly or largely of glass)	Vitrophyre (containing phenocrysts)			
	Obsidian and Pitchstone (without phenocrysts)			

[1] "Orthoclase" includes microcline, sanidine, and microperthite. B = biotite.
may be

the feldspar can be determined only by optical means. However, it is a distinction that should always be attempted in the field. The principle according to which the felsophyres are subdivided is that they normally carry as phenocrysts the essential minerals that make up the coarse-grained rocks (phanerites) of which they are the chemical equivalents.

of Common Igenous Rocks

Classes Based on Mineral Composition

Medium-Colored Rocks		*Generally Dark-Colored Rocks*	
Plagioclase, Chief Feldspar		Plagioclase, Chief Feldspar	No Feldspar
B H P	H B P	P H O	O P H B
with quartz	without quartz	without quartz	No quartz
Quartz diorite (Tonalite)	Diorite	Gabbro	Peridotite Hornblendite Pyroxenite
Micro-quartz-diorite (Microtonalite)	Microdiorite	Microgabbro (Dolerite)	
Quartz diorite porphyry (Tonalite porphyry)	Diorite porphyry	Gabbro porphyry	
Dacite	Andesite	Basalt	
Felsite and Felsophyre			
Vitrophyre (containing phenocrysts)		Basalt glass (Tachylyte)	
Obsidian and Pitchstone (without phenocrysts)			

H = hornblende. P = pyroxene. O = olivine. One or more of these minerals present.

Subdivisions of glasses. The rocks of class IV, glasses, are those consisting wholly or partly of glass; they are distinguished by their glassy or resinous luster and want of stony texture. They are classified as follows:

A. Obsidian: luster strong, bright, glassy; color generally black, but also red, more rarely brown and greenish.

B. Pitchstone: luster pitchlike or resinous; colors various, as above, but black is less common.

C. Perlite: glassy rock having perlitic structure resulting from small spheroidal fractures produced by concentric shrinkage cracking; generally gray.

D. Pumice: highly vesicular glass (see page 129); generally white or very light colored.

All these rocks range from nonporphyritic to porphyritic, but porphyritic fabric is much less common in pumice than in the other three. If they are porphyritic, a general name for them is *vitrophyre* (glass porphyry). Several varieties can be distinguished, as in the porphyries of the class above, according to the phenocrysts that predominate; thus *quartz vitrophyre*, *feldspar vitrophyre*, etc. Most of the rocks of this class have originated as lavas.

PYROCLASTIC ROCKS. In addition to the three main classes of igneous rocks described above, some authorities add a fourth class, the rocks formed from fragmental material blown out during volcanic eruptions, already mentioned on page 112 as *tuffs* and *breccias*. They are collectively termed pyroclastic rocks.

Such material serves as a connecting link between the sedimentary and the igneous rocks. For, as the volcanic ejecta falls through the air, it becomes assorted as to size, and successive ejections will thus produce a rough but distinct bedding. Or it may fall into water and become perfectly stratified. Falling on the land, it may bury vegetation and consequently contain fossil imprints of plants and leaves; or, if it falls into water, it may contain the remains of aquatic organisms. Hence if we classify pyroclastic rocks as igneous rocks, we cannot say that a distinguishing feature of igneous rocks is that they do not contain fossils. See remarks on page 103.

CLASSIFICATION TABULATED. The classification adopted and described in the foregoing pages is shown for convenience of reference in Table 1.

Classifications based on microscopic research. In the classification previously described, the color and texture of rocks play a prominent part, and mineral composition can be used only in an approximate manner. But if rocks are studied in thin section under the petrographic microscope, all the minerals and their exact characters can be discovered and their relative proportions determined. In this exact work the kinds of rocks that are recognized by petrographers are based primarily on the kinds, and to some extent on the relative proportions, of the component minerals. This work has led to the recognition of about 800 kinds of igneous rocks that have been named. Generally the rocks are grouped, first, according to mineral composition, and second, according to texture. Some petrographers lay weight also on the mode of occurrence of the rocks, whether extrusive or intrusive, while others add to this the genetic relations or groupings that the rocks have in nature. Classifications have also been proposed in which the chemical composition of the rocks plays the most prominent part.

CHAPTER VIII

DESCRIPTION OF IGNEOUS ROCKS

PHANERITES

As explained in the chapter on classification, phanerites are those igneous rocks made up of mineral grains large enough to be identified with the unaided eye or lens. Rocks whose constituent grains are too minute to be identified in this way are placed under the heading of aphanitic igneous rocks.

GRANITE

Composition. Granites are equigranular phanerites composed of feldspars and quartz. Some consist wholly of these minerals, but generally biotite is present; hornblende rarely occurs.

Feldspar predominates and is readily recognized by its appearance and cleavage. Only one kind of feldspar is present in some granites, but generally two kinds occur: orthoclase and, subordinately, plagioclase. They can be distinguished apart in some rocks by their color: if one feldspar is flesh colored, pink, or red and the other is white, gray, or yellow, pretty certainly the first is orthoclase, and the second is plagioclase. Basal cleavage surfaces of the plagioclase on close inspection may show the characteristic twinning striae (see page 129), but if the plagioclase grains are too small the striae cannot be seen.

Granites containing about as much plagioclase as orthoclase are called *quartz monzonites,* according to a usage widely current in recent years. Those in which the plagioclase exceeds the orthoclase in amount are called *granodiorites*; but the distinction between quartz monzonite and granodiorite can generally not be made by megascopic examination and consequently they are classed as granites.

Quartz is normally anhedral, filling the interstices between the other minerals, and hence it does not have its own characteristic crystal shape. It is generally colorless, white to dark, or smoky gray; some is red from included hematite; more rarely it is bluish. Quartz is recognized by its glassiness, hardness (7), lack of cleavage, and conchoidal fracture.

Mica may be present either as black biotite or as silvery muscovite, or both may be present. Muscovite is rarely the only mica. Hornblende occurs in black or dark-green grains or prisms. In some granites it is the only dark mineral, but it is generally accompanied by biotite.

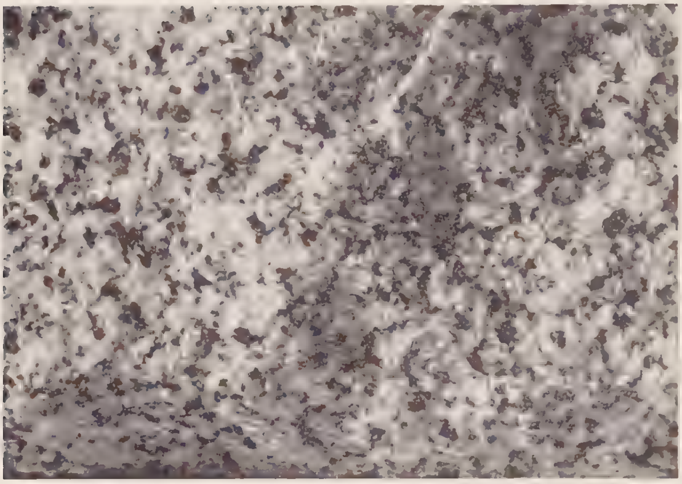
Feldspar, quartz, and biotite, named in descending order of abundance, are the chief minerals of the average granite. If a granite is fairly coarse grained, close inspection will commonly show sporadic metallic-looking grains of magnetite. Other minerals may be seen—honey-yellow crystals of sphene, dark-red grains of garnet, etc.—but these are minor constituents and are of no consequence in determining the rock.

Texture. In ordinary normal granite the fabric is equigranular, that is, the feldspar, the quartz, and the biotite roughly (very roughly in some granites!) approximate one another in grain size. In the same mass, the texture may grade insensibly from this type into that of gneiss, which becomes noticeable through the parallel arrangement of the components, especially the micas. Thus the normal granite has graded into granite gneiss of the kind known as primary, or fluxion gneiss. The texture of normal granite is shown on Plate 13. Some granites contain orthoclase crystals that are larger than the average grain and have distinct crystal form. Such granite is the porphyritic granite described below. The dark minerals in some granites tend to group or bunch together in spots.

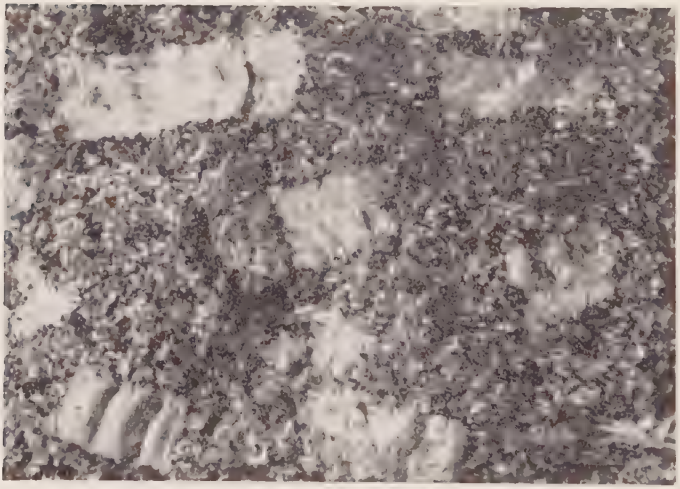
Color. The general color of granites depends largely on that of the feldspar and on the proportion of this feldspar to the dark minerals. Thus the color grades from white to gray to dark gray, as the result of the presence of the biotite or other dark minerals. Such types are very common wherever granites are abundant. More rarely the quartz and feldspar are themselves gray or dark gray and thus determine the color. An example of this is the granite of Quincy, Massachusetts, largely used as a building stone. Another common type of coloring is one in which the granite is flesh colored, pink to red, or even deep red. Such red granites occur in many localities, for example, in Colorado where they make up the mass of Pikes Peak. They are quarried and much used for building.

Varieties. The varieties of granite depend on the relative proportions of the light and dark minerals, the color, and the texture. The relative amount of biotite (or hornblende) to quartz and feldspar has a wide range; biotite or hornblende may fail entirely, or they may occur abundantly and give the granite a dark color. The grain size may exceed 2 centimeters. These variations, along with those in color, produce distinct kinds of granite of marked individualities, many of which have received local names.

PLATE 13.



A. NORMAL GRANITE.



B. PORPHYRITIC GRANITE.

Porphyritic granite. As mentioned above, the feldspar may occur partly in large distinct crystals, or phenocrysts. Crystals as large as two and three inches in diameter may be scattered through the granite, but because these phenocrysts are as a rule rather sparsely distributed, such a granite is customarily termed a *porphyritic granite* rather than granite porphyry. The feldspar phenocrysts are potassium feldspar (orthoclase or microcline) and have the forms shown under feldspar, page 27. Reflection of light from the cleavages of these phenocrysts generally shows them to be in twin halves, due to Carlsbad twinning. A porphyritic granite of this kind is illustrated in Plate 13.

Chemical composition. The bulk compositions of a few granites are shown in Table 1 to illustrate the kind of magma from which such rocks have formed. The large percentages of silica, alumina, and alkalies in the analyses of the granites explain the strong predominance of feldspars and quartz in these rocks. The larger amount of lime in the analysis of the granodiorite than in those of the granites causes plagioclase feldspar to dominate over orthoclase in the granodiorite. Furthermore, the increased iron and magnesia indicate a larger content of dark minerals. Coincidentally with these increases the silica content is smaller.

TABLE 1

ANALYSES OF GRANITES AND GRANODIORITE

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O^+	XyO[1]	Total	Sp. Gr.
I...	77.0	12.0	0.8	0.9	0.0	0.8	3.2	4.9	0.3	0.1	100.3	....
II...	71.6	14.5	1.5	1.1	0.9	2.0	3.0	4.1	0.8	0.5	100.0	2.63
III...	64.0	15.6	2.0	2.8	2.2	4.2	2.8	4.1	0.7	0.9	99.3	2.72

[1] XyO represents small quantities and traces of other oxides.

I. Granite, Sentinel Point, Pikes Peak, Colorado.

II. Average biotite granite, average of 90 analyses (Tschirwinsky).

III. Granodiorite, average of 4 closely agreeing analyses, Butte, Montana (consists of 42 per cent of sodic plagioclase (An_{40}), 18 per cent of orthoclase, 23 per cent of quartz, 10 per cent of biotite, 4 per cent of hornblende, and 3 per cent of minor components).

Physical properties. Granites range in specific gravity in accordance with the kinds and relative amounts of their component minerals. From 2.61 to 2.75 is the ordinary range, the heaviest being those richest in ferromagnesian minerals. Granite of specific gravity 2.64 averages 165 pounds to the cubic foot.

The porosity of granites quarried for building purposes is generally very small, the percentage of water absorbed, compared with the weight of the dry rock, being about 0.15 per cent. Thus a cubic foot of average granite if completely saturated contains 4 ounces of water. The strength of granites in resistance to crushing is very great, as a rule far greater than any load they would be called upon to bear in architectural work. A series of Wisconsin granites, for example, has crushing strengths ranging from 15,000 to 40,000 pounds per square inch; from 15,000 to 20,000 is perhaps the average. If granite had been used in the construction of the foundation of the Washington Monument, which has at its base a pressure of 342.4 pounds per square inch, it would have had a great excess of strength, even under the load of this extraordinarily high stone structure.

Uses of granite. As is well known, granite because of its great strength and durability is extensively used for architectural purposes. Its pleasing colors and the high polish it takes cause it to be employed as an ornamental stone in interior work, in monuments, etc. In one respect, however, many granites have a defect that somewhat impairs their value for use in buildings in large cities. This defect is that they do not resist fire well, but crack, scale, and crumble under great heat. Probably the large coefficient of expansion of quartz at high temperatures is the chief cause of the poor fire-resistive capacity of granite. Another cause is its poor thermal conductivity.

Jointing in granite. Granite tends to have block jointing on a large scale. Generally there are three distinct sets of joints, two of which are approximately vertical; the third is horizontal. These sets may be nearly at right angles, producing cubes, but more commonly they are at angles that make rhomboidal blocks. For some masses the horizontal jointing is the most pronounced, and such masses have a sheeted or layer-like character suggesting bedding. In dikes the joints are much more numerous and the dikes break into small blocks and plates.

The jointing of granite is of much importance in the work of excavation in mining, tunneling, and quarrying because it facilitates the removal of material. It explains also why every granite mass is not suited to furnish material in blocks large enough for constructional purposes. Quarries such as those in the Rapakivi granite in Finland, from which the base of the Alexander monument in Leningrad, a cube of 30 feet and its shaft, 100 feet high by 15 feet in diameter, were taken, and such as those in Egypt from which the great obelisks were cut, are not common.

Erosion forms of granite. Jointing largely determines the work of erosive agents on the granite masses, but the topographic forms produced by erosion depend also on the severity with which these agents act. In the high mountain chains and wherever these agents are energetic, spires, needles, and castle-like forms are produced; but in the deeply eroded ranges and where glaciation has been pronounced, the granite masses form more smoothly modeled, rounded, or dome-shaped bodies with gentle slopes and broad valleys, such as are seen in the hills and mountains of New England and the southern Atlantic States. The views on Plate 14 illustrate the contrast in the erosion of high mountain zones and low, worn-down regions.

Orbicular granite. The component minerals of some granites, instead of consisting uniformly of grains of about the same size, are aggregated in spots in an unusual way and arranged in ovoid or spherical bodies. Thus a granite from Craftsbury, Vermont, called "pudding granite," is full of nodules, ranging from the size of a pea to that of a nut, composed almost entirely of leaves of black mica, as shown on Plate 15. More commonly the nodules are composed of several minerals and consist of a nucleus surrounded by one or more shells. The component minerals are the same as those in the main body of the rock, but their proportions differ in the nucleus and in the shells. Some shells consist mostly or entirely of sialic minerals, while other shells consist mostly of ferromagnesian ones. Their appearance is shown on Plate 15.

The bodies are round, ovoid, or spindle-shaped, as if they had been drawn out. Formerly they were thought to represent pebbles and were considered to be proof of the metamorphic origin of granite from conglomerates, but the arrangement and regular internal structure of the ovoids precludes such an idea. They are now generally held to be due to some process of aggregation of material in the magma, though some of them may represent inclusions of other rocks (xenoliths) that have been modified by the action of the enveloping magma. Granites of this kind are called *orbicular*; they are not common. Similar orbicules occur also in diorites and gabbros.

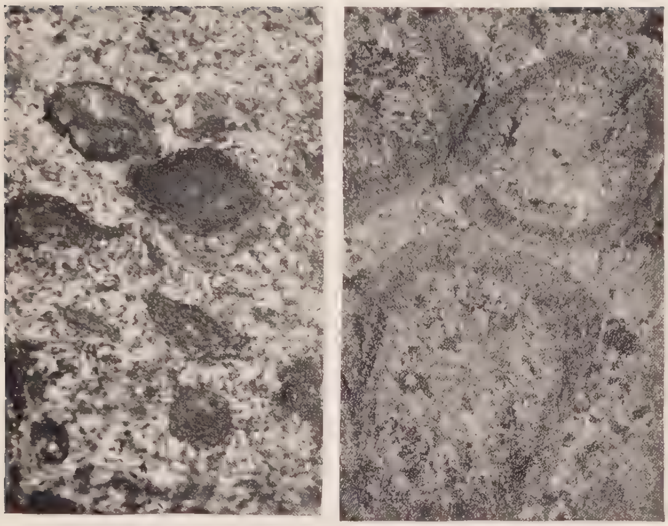
Miarolitic structure. Most granites show little or nothing of the miarolitic structure described on page 131. The conditions under which most granites were formed, presumably at considerable depth, were therefore unfavorable for forming such cavities. In other occurrences of granite, presumably formed at higher levels in the Earth's crust, miarolitic cavities occur, and handsomely crystallized minerals can be seen on their drusy surfaces. The crystals from miarolitic cavities in the granite of the Pikes Peak region, Colorado, from the Mourne

PLATE 14.

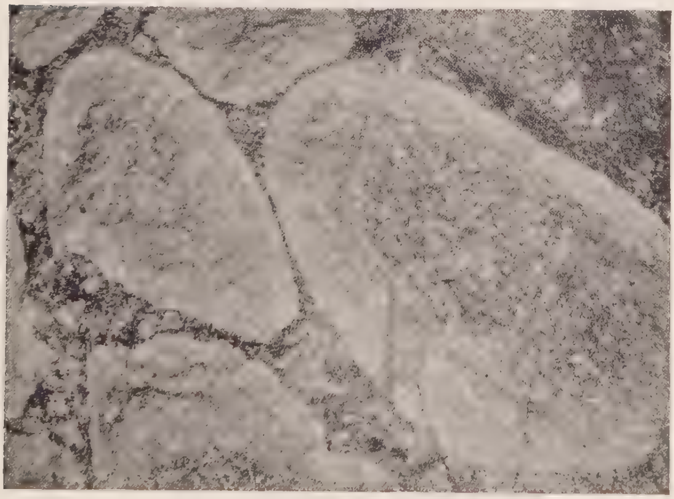
A. EROSIONAL FORMS OF GRANITE IN THE HIGH ALPS.
(After Duparc.)

B. EROSIONAL FORMS OF GRANITE IN OLD AND LOW MOUNTAIN REGIONS, STONE MOUNTAIN, GEORGIA.
(Georgia State Geological Survey.)

PLATE 15.



A. Craftsbury, Vermont. B. Kortfors, Sweden.



C. Stockholm, Sweden.

ORBICULAR GRANITES.

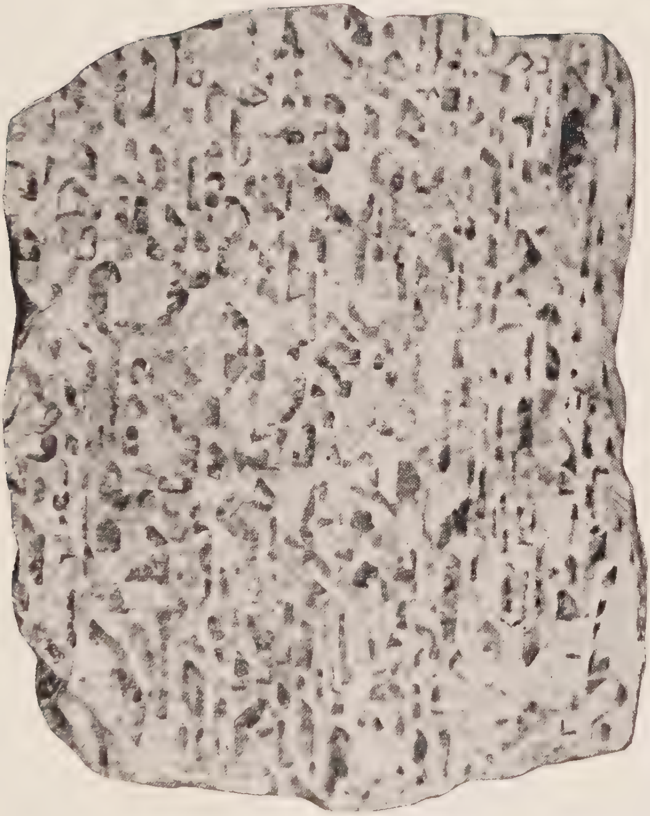
Mountains, Ireland, from Baveno on Lago Maggiore in northern Italy, and from other localities are well known in mineral collections.

Pegmatite. Pegmatite is abundant as dikes and veins in many granite masses and in the adjacent country rock. So common is pegmatite of this kind that when the term pegmatite is used granite pegmatite is understood to be meant, unless it is otherwise specified. Pegmatites have the general characters described on pages 145 to 148. The chief minerals are quartz and feldspar, the latter being mostly the potassium feldspar *microcline*. Albite also occurs, generally as a replacement product formed at the expense of the microcline. The quartz and microcline tend to be intergrown in such a manner that the quartz individuals, which are long rods, show figures that recall the script used in Arabic writings when seen in cross section on cleavage surfaces of the microcline. Because of this resemblance the intergrowth is called *graphic granite* (see Plate 16). It indicates that the quartz and microcline have crystallized simultaneously.

In most pegmatites, however, the minerals occur separately, in some as huge crystals, measurable in feet or yards. Pegmatites are therefore mined commercially, the feldspar being used in the manufacture of china, porcelain ware, and other ceramic products. Large crystals of muscovite mica occur in the pegmatites of some localities. Such mica-bearing pegmatites are the sole source of this valuable material, which is so much used as an insulator in electrical machinery and radio equipment. In addition to these chief minerals, many others occur, the more common of which are tourmaline, garnet, beryl, and spodumene among the silicates, apatite and other phosphates, and various minerals containing rare earths. Some of these minerals—the colored tourmalines, topaz, and beryl—are valuable because they afford material suitable for cutting into gems or semi-precious stones; others are useful as sources of the rarer elements used in chemical industry. Uraninite, which is restricted in its occurrence chiefly to granite pegmatites, is of great interest because it enables us to determine the age of the pegmatites in years by means of radioactive methods. Pegmatites, however, are not commercial sources of uraninite; at a few places in the world quartz veins are sources. A full list of all the minerals known to occur in granite pegmatites would include a large proportion of all the kinds known in mineralogical science.

Inclusions. Schlieren. Granites generally contain inclusions of the various kinds described on pages 137 to 138. Some inclusions show by their composition and angular shapes that they are fragments of previously existing rock formations that were broken off and engulfed in the granite magma. Inclusions of this kind (xenoliths) are most

PLATE 16.



GRAPHIC GRANITE.

abundant near the borders of intrusive granite masses. They range in size from less than an inch long to many yards long. When they occur in the center of an area of granite, it can be suspected that the granite has not been eroded deeply below the roof of other rocks that formerly covered it.

Some of what appear to be inclusions are the *schlieren* previously described. Some schlieren consist wholly of quartz and feldspar, or nearly so, and thus are light in color; some are rich in biotite, hornblende, or both and therefore dark in color. Such dark streaks may have been xenoliths that were smeared out by viscous drag of the magma, but some may have been formed by aggregation of the normal dark minerals of the granite while they were crystallizing from the magma.

Complementary dikes. Many granite masses are cut by complementary dikes, as described on page 140. The *leucocratic* dikes are composed almost wholly of quartz and feldspar. The rock of the dikes has a finely granular, sugar-textured appearance. This variety of granite is called *aplite.* Small black specks of biotite or hornblende or of black tourmaline may be seen in it, but dark minerals invariably play a very minor role. Aplites are typically very light in color: white, flesh color, pale yellow, or gray. Some aplites are so fine grained that they grade into felsites of the colors mentioned, and some are porphyritic with phenocrysts of quartz or feldspar or both and are consequently granite porphyry or felsite porphyry. Most commonly, however, they are equigranular with a grain size about like that of cube sugar. They characteristically occur as dikes. The dikes are inclosed mostly within the parent granite masses; less commonly they occur in the surrounding rocks. They range in thickness from a fraction of an inch to many yards, but the thinner dikes are prevalent. Some of the larger dikes when traced along the outcrop are found to grade into pegmatite.

The *melanocratic* dikes and sills, complementary to the quartzo-feldspathic aplites just described, are dark or black heavy rocks of basaltic aspect. They are composed chiefly of biotite, hornblende, and pyroxene, together with feldspars, but the grain size is commonly too fine for these minerals to be distinguished and, consequently, the rocks are likely to be identified as basalts. Technically, they are lamprophyres. Many of them in which biotite is prominent, have a characteristic glimmering appearance, and in others the plates of biotite form conspicuous phenocrysts; these rocks are known as *mica traps.* Porphyritic crystals of hornblende, as well as of augite, may occur in them. Not uncommonly they enclose xenoliths of granite and other rocks. They weather readily into soft greenish material or into brown earthy masses. The earlier stages of weathering are marked by the formation of carbonates, and such altered rocks effervesce freely when treated with acid.

The lamprophyres occur characteristically in narrow dikes, generally but a few feet thick, and also in sills and small laccoliths. The origin of these complementary dikes was discussed on pages 140 to 141.

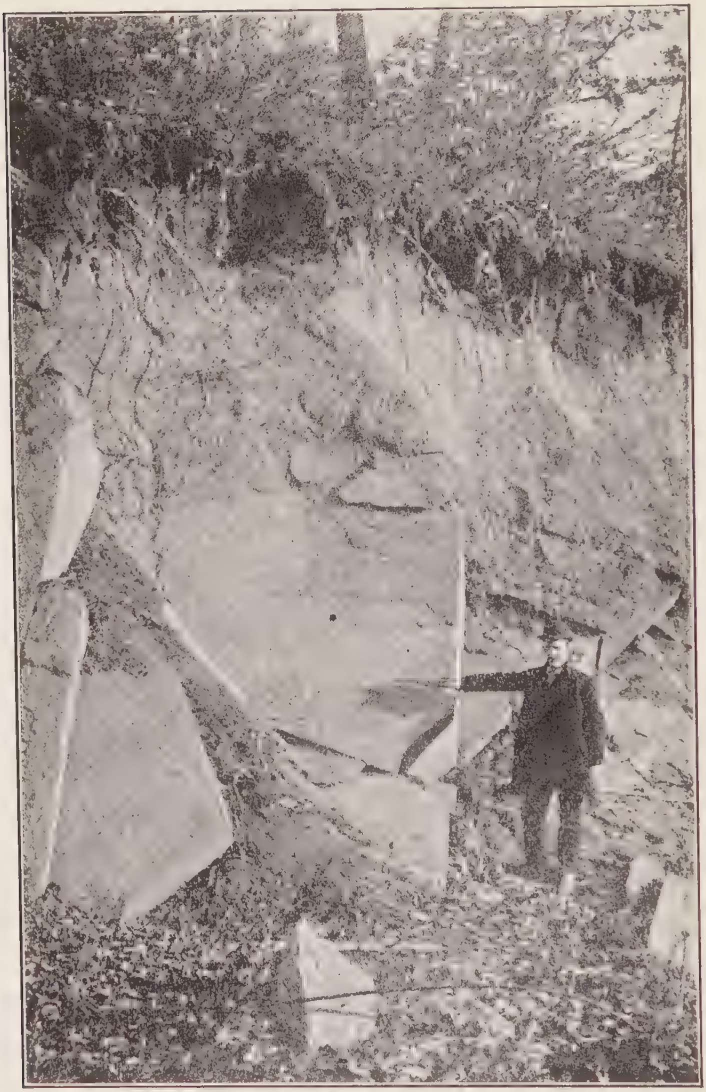
Contact phenomena. Around large granite intrusive masses the contact phenomena described on page 149 attain their greatest development and perfection. As a result of contact chilling the granite may become a felsite or granite porphyry in the border zone, or it may have a differentiated border zone (see page 139) and become so enriched in dark silicates as to pass into a diorite border facies. More rarely, on the other hand, the border zone is so poor in dark minerals as to present a border facies of aplite, quite like that seen in the complementary dikes. The chilled border zones are purely textural modifications; the border facies are mineralogical. More rarely, granites have a border facies of pegmatite.

The external contact-metamorphic changes described in the previous chapter are far more common and have been more extensively studied in connection with granites than with any other variety of igneous rock, as a result of the widespread occurrence of granites. Around great granite batholiths the contact-metamorphic effects are profound and far reaching, involving tracts as much as several miles wide. Such tracts are of great interest and importance, not only from the geologic standpoint, but also because many of them are the sites of valuable ore deposits. If granite comes directly against sedimentary rocks with vertical contact and the sedimentary rocks show no evidence of metamorphism, it can be safely inferred that faulting or dislocation has brought them together.

Weathering of granite into soil. Through the action of the atmosphere, of water, and of heat and cold, granite breaks down into soil. In temperate regions, such as eastern North America, the initial change is largely a mechanical disintegration and the resultant material differs in chemical composition but slightly from the original rock.

As it changes, however, granite alters considerably in appearance. The biotite tends to bleach and lighten in color, and ferrous compounds tend to become ferric hydrate and migrate, staining the rock yellowish red to dark brown. At the same time the granite loses its firm texture and becomes more or less crumbly. Finally it falls into

PLATE 17.



GNEISSIC GRANITE, GRADING FROM FIRM ROCK BELOW INTO WEATHERED ROCK AND SOIL ABOVE; ROCK CREEK PARK, WASHINGTON, DISTRICT OF COLUMBIA.

(G. P. Merrill, *Bulletin*, Geol. Soc. America, vol. 6, p. 321, 1896.)

an angular sand, composed mainly of particles of quartz and feldspar. See Plate 17.

From this stage, as the change into soil becomes more pronounced, the most important process is the conversion of the feldspar into kaolinite, according to the following reaction:

$$\overset{\text{Orthoclase}}{2KAlSi_3O_8} + \overset{\text{Water}}{2H_2O} + \overset{\text{Carbon dioxide}}{CO_2} = \overset{\text{Kaolinite}}{H_4Al_2Si_2O_9} + \overset{\text{Quartz}}{4SiO_2} + \overset{\substack{\text{Potassium}\\\text{carbonate}}}{K_2CO_3}$$

This reaction begins as soon as the granite is attacked by downward percolating water. It has no sooner started than the rock begins to crumble, but after that it goes on slowly but steadily until the feldspar is eventually changed into kaolinite. As a result, a soil consisting of a mixture of kaolinite and quartz sand, stained reddish or yellowish by iron compounds, is formed. Such a soil is called *loam*. Generally the process has not gone entirely to completion and the soil contains more or less feldspar in small particles, undergoing alteration. This residual feldspar has an important bearing on the maintenance of the fertility of the soil.

The soil-forming process is one of the most important that takes place in the great laboratory of Nature, for by means of it not only is solid rock converted into soil, but one of the most essential of plant foods, potash, is converted into a soluble form in which it can be utilized by plants. At the same time the other essentials of plant food, such as silica, magnesia, and lime, also are unlocked from the rocks and rendered available. By the aid of these mineral foods plants are able to grow and to synthesize from water and carbon dioxide those substances on which all animal life ultimately depends.

In certain tropical regions having a climate characterized by a short wet season alternating with a long dry season, the decay of granite gives rise to a red or yellow-brown ferruginous earth called *laterite*. It consists of a mixture of quartz sand and gibbsite, $Al(OH)_3$, colored by iron hydrates. During lateritization the feldspar of the granite alters to gibbsite, and even kaolinite alters to gibbsite and allied aluminum hydrates. Lateritization therefore involves the break-down of silicates and the removal of their silica from the rock. Apparently under conditions of intense lateritization even quartz is eliminated, but this has not yet been proved. Laterites derived from quartz-free rocks do not contain quartz, of course, and such laterites also are widely prevalent.

The surface of granites and other rocks in hot deserts is coated with a brownish or black skin, having a lusterlike varnish and therefore appropriately called *desert varnish*. It is supposedly the result of the alteration of the iron-bearing components and the formation of iron and manganese compounds.

Occurrence of granites. Granite is the most common and widely occurring igneous rock and plays a prominent role in the foundation of the continental masses. As stocks and batholiths it makes the central core of many of the great mountain ranges, where it has been revealed by later erosion. In those parts of the Earth that have been subjected to repeated disturbances of the crust and have later been uncovered by erosion, granites are common. Thus great masses of granite, intrusive at different ages, occur in eastern Canada, in New England, and generally in the Piedmont province from southern New York to Georgia. They occur in Missouri, Wisconsin, and Minnesota, but, in general, the Central States, which compose the Mississippi Valley, being covered with undisturbed stratified rocks, are devoid of intrusive igneous masses, though in recent years, as a result of drilling for oil, granites have been shown to form a large part of the basement on which the later sedimentary rocks lie. In the Rocky Mountains and in the Far Western States granites are of great importance, not only because of their vast areal extent but also because they were the "ore-bringers" in most of the mining districts of that richly endowed metalliferous region. Likewise in Europe, Africa, Asia, and Australia they appear frequently and are exposed over vast areas. A list of occurrences can be almost indefinitely lengthened, but enough has been said to show the extent and world-wide distribution of granite.

SYENITE

Composition. Syenites are phanerites composed chiefly of orthoclase. They differ from granites in containing no quartz or only a negligible quantity. They may consist entirely of feldspar, but generally some hornblende, biotite, pyroxene, or combination of these is present. The dark minerals are subordinate in amount to the feldspar, however; in the normal syenite they make up $\frac{1}{8}$ to $\frac{3}{8}$ of the bulk composition.

According to which of the ferromagnesian minerals predominates, the rock is spoken of as hornblende syenite, biotite syenite, or augite syenite. All these varieties are here treated under the general heading of syenite. There is one variety of syenite, however, that has a particular mineral composition that makes it of special interest. This variety can generally be identified by observing carefully the special features that it presents; otherwise it cannot be distinguished and will be classed in the general group of syenites. The syenite of this kind contains in addition to the feldspars and other minerals a notable amount of nepheline, or of this mineral and its congener, sodalite. Syenites are divided as follows:

A. Syenite, in general consisting chiefly of orthoclase, without quartz.

B. Nepheline Syenite, consisting chiefly of alkali feldspars, with nepheline.

Properties of syenite. Syenites are generally equigranular. In some, however, the feldspar tends to have a tabular form like that of a book; its cross sections on the rock surface are therefore lathlike. In some syenites the tabular feldspars are in more or less parallel position, an arrangement probably brought about by flowage movement of the fluid mass during crystallization. Porphyritic varieties also occur as in granite, and these grade into syenite porphyry. The color is varied as in granites: white to pink or red, and gray or yellow tones are common, gray especially so. The specific gravity varies with the minerals and their proportions; it may extend from 2.6 to 2.8. In a tendency to miarolitic structure, in jointing, in erosional forms, in alteration into soil, in inclusions, and in contact metamorphism, what has been said in regard to granite applies also to syenite and need not be repeated.

Syenite masses are also accompanied by pegmatite dikes, but these are not so common as the granite pegmatites. Syenite pegmatites also yield a great variety of minerals.

Chemical composition. Chemically, syenites are distinguished from granites by their small content of silica, which accounts for the absence of quartz; in other respects they resemble granites. The chemical characters are shown in Table 2.

TABLE 2

ANALYSES OF SYENITES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
I...	60.7	19.6	1.5	3.0	0.8	2.3	4.9	5.9	0.3	0.6	99.6
II...	60.2	20.4	1.7	1.9	1.0	2.0	6.3	6.1	0.3	0.4	100.3
III...	61.6	15.1	2.0	2.2	3.7	4.6	4.3	4.5	0.7	1.0	99.7
IV...	62.5	16.5	2.4	2.0	1.9	4.2	4.4	4.6	0.6	1.3	100.4

[1] XyO represents small amounts of other oxides.

I. Belknap Mountains, New Hampshire.
II. Fourche Mountains, Arkansas.
III. Little Belt Mountains, Montana.
IV. Plauen, near Dresden, Germany (specific gravity 2.73).

Occurrence of syenites. Syenites are not common rocks. Some occur in independent masses, but others are connected with masses of granite, which as the result of diminishing content of quartz grades into syenite; a syenite body of this kind is a facies of the larger mass of granite. In the United States syenite occurs most abundantly in the Adirondacks. In comparison with the great batholiths of granite distributed so generally in the continental masses, syenites are of relatively small volume.

Uses of syenite. For all constructional and other commercial uses syenite has the same value as granite. Because of its relative rarity compared with granite, however, it is little used. Its crushing strength is equal to that of granite. Its weight per cubic foot is about the same. The absence of quartz, a mineral harder than feldspar, should make it an easier stone to dress and polish, and practically it resists weathering as well, if not better. Absence of quartz makes it also a better stone in resisting the heat of fires (compare granite, page 170) and in consequence it would be a more advantageous material to use for building in our large cities. If these advantages over granite were more generally understood, the accessible occurrences in New England would probably be more extensively exploited. The beautiful dark gray syenite of south Norway, with its striking opalescent feldspars, is much used in northern Europe as an ornamental stone.

Nepheline syenite. Nepheline syenite is distinguished by the fact that in addition to the feldspars, which are almost wholly alkalic in composition, much nepheline is present. Some nepheline is flesh colored, but it is generally smoky gray. In its lack of good cleavage it resembles quartz, but it can be distinguished from quartz by the test for gelatinization (page 90). It generally occurs as anhedral grains mixed with the feldspars, but some may show imperfect crystal form, namely, short stout hexagonal prisms. It is likely to be accompanied by sodalite, which is generally in bright-blue grains or, if the rock is coarse grained, in lumps and masses; if sodalite is present it is useful in aiding to distinguish nepheline syenite from common syenite. Noselite and cancrinite may also be present. Biotite (lepidomelane), hornblende (arfvedsonite), and pyroxene (aegirine) are usually present in variable amounts, in black plates, grains, or prisms and containing considerable sodium and iron. The presence of sodium in the minerals of nepheline syenite is readily understood from a consideration of the chemical analyses, given in Table 3, which show the composition of the magma from which they crystallized.

The *color* of nepheline syenites is varied but is commonly gray. The *texture* is equigranular. The tabular habit of the feldspars mentioned above is common. Nepheline syenites are likely to contain many accessory minerals but usually only in microscopic sizes; some of these are of especial interest on account of the rare earths they contain. In places great variations of the constituent minerals give rise to widely differing facies. Many of these varieties have received special names.

Nepheline syenite is commonly cut by complementary dikes, and these differ in composition from those associated with granites and common syenites. One is a pale-brown or pink felsite, another a bright- to dark-green rock called *tinguaite*, which owes its green color to microscopic needles of aegirine. Tinguaite generally contains nepheline and gelatinizes with acid. The lamprophyres complementary to these are heavy dark rocks of basaltic aspect called camptonite, which show abundant phenocrysts of biotite, augite, or hornblende.

The chemical composition of nepheline syenites is illustrated in Table 3. The most striking thing in the composition of these rocks is the high amount of alumina and alkalies, with moderate silica. This relative deficiency of silica has caused nepheline ($Na_2O{\cdot}Al_2O_3{\cdot}2SiO_2$) to form rather than albite ($Na_2O{\cdot}Al_2O_3{\cdot}6SiO_2$), there being not enough silica to convert all the alumina and alkalies into feldspar.

TABLE 3

ANALYSES OF NEPHELINE SYENITES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
I..	58.8	22.5	1.5	1.0	0.2	0.7	9.6	4.9	1.0	0.3	100.5
II..	53.1	21.2	1.9	2.0	0.3	3.3	6.9	8.4	1.4	2.0	100.5

[1] XyO represents small quantities of various oxides.

I. Salem Neck, Massachusetts.
II. Magnet Cove, Arkansas.

From this it can be inferred that quartz and nepheline cannot crystallize from the same magma; the excess silica represented by the quartz would convert the nepheline into albite, and therefore quartz and nepheline do not occur in the same rock. Some noselite may be present; cancrinite may also occur, and of the associated minerals zircon is perhaps the most characteristic. In this connection the description of the feldspathoid group in the part dealing with minerals should be read.

Pegmatite dikes occur in genetic connection with nepheline syenites. Those of south Norway and Greenland are especially interesting for their great variety of minerals, many of which are composed in part of the rarer elements.

Nepheline syenites occur as a rule in rather small masses; relatively large bodies of them are known to occur in only a few places: Greenland; south Norway; Pilansberg, South Africa; and Kola Peninsula, Karelia, Russia. They are uncommon and are geologically of small importance compared with granites, gabbros, and diorites. In the United States they are found in New England and in Arkansas, where they are the parent rocks of the bauxite deposits.

Nepheline syenite is too uncommon to be commercially important as a building stone, but it is used where it occurs, making an excellent stone. In the neighborhood of Magnet Cove, Arkansas, it has been thus used. In recent years it occurred to the Russians to use nepheline syenite as a ceramic raw material; this idea has generally been adopted, with the result that our domestic producers of feldspar are experiencing stiff competition and would like to have a tariff put on Canadian nepheline syenite.

Corundum syenite. Among the different varieties of syenite described above, some contain, in addition to the normal constituents, a notable amount of corundum. That this mineral was formed is due to the fact that the magma contained more alumina than the alkalies and lime present could turn into feldspars and feldspathoids, and the excess alumina crystallized out as corundum (Al_2O_3), just as in granites the excess of silica forms quartz (SiO_2). The corundum occurs in crystals, either hexagonal prisms or barrel-shaped, in grains, and in lumps; it is generally gray. It is easily identified by its extreme hardness.

Corundum has been found in syenite and nepheline syenite in Renfrew, Hastings, and other counties of Ontario, Canada, where these rocks occur in a belt more than a hundred miles long; in the Ural Mountains; in Coimbatore district, India; and, especially, in South Africa. Some of these occurrences, notably those in South Africa, are of economic value as a source of this valuable abrasive.

DIORITE

Composition. Diorites are phanerites composed of plagioclase feldspar and one or more ferromagnesian minerals. The amount of dark minerals in the normal diorite is between one-eighth and three-eighths. The dark minerals are biotite, hornblende, or pyroxene, occurring either singly or together. The hornblende is generally black but may be dark green. Although some is in bladed or prismatic forms, it is commonly in short thick crystals or grains. At best it is difficult, and generally it is impossible, to distinguish the pyroxene from the hornblende. The plagioclase present in diorites as defined petrographically is generally a sodic variety (andesine). This determination cannot of course be made on the hand specimens, because the composition of the plagioclase feldspars cannot be determined megascopically.

Quartz is likely to be present and can generally be identified by means of the lens; if it occurs in notable quantity (more than 10 per cent) the rock is called quartz diorite. In recent years the name *tonalite* has been revived as a synonym for quartz diorite and become widely used.

Although the rocks determined as diorites by this megascopic classification correspond in a general way with most of the diorites of the more strict classifications founded on microscopic and chemical methods, they include some gabbros and less common rocks.

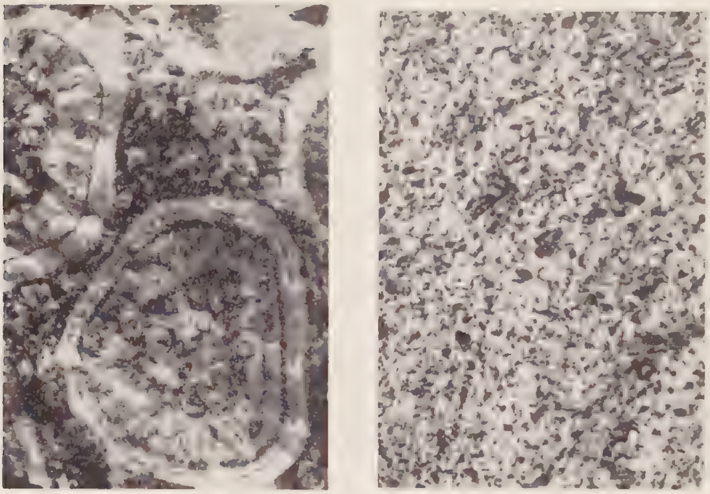
General properties. The color of diorites is gray to dark gray or greenish. It is determined by the proportion of dark minerals to feldspar. The different varieties of diorite are determined by color, coarseness of grain, and mineral composition. The fabric is equigranular. Porphyritic fabric, although not unknown, is far less common than in granite. The black hornblende prisms in some diorites are large enough to produce an impression of porphyritic texture, but this impression is dispelled as soon as one compares it with the average size of the surrounding mineral grains. Orbicular structures are known to occur. An orbicular diorite from Corsica has been used to some extent as an ornamental stone; it is illustrated in Plate 18. Miarolitic cavities occur as in granite but are commonly masked by having become filled with calcite. Fluidal or roughly parallel arrangement of the component minerals may occur, giving the diorite a more or less pronounced gneissic structure.

Diorite pegmatite dikes also occur. Their composition is like that of the diorite with which they are associated, except that they are somewhat richer in the light-colored minerals. Diorites are cut by complementary dikes much like those in granites.

The jointing of diorites is like that described for granites.

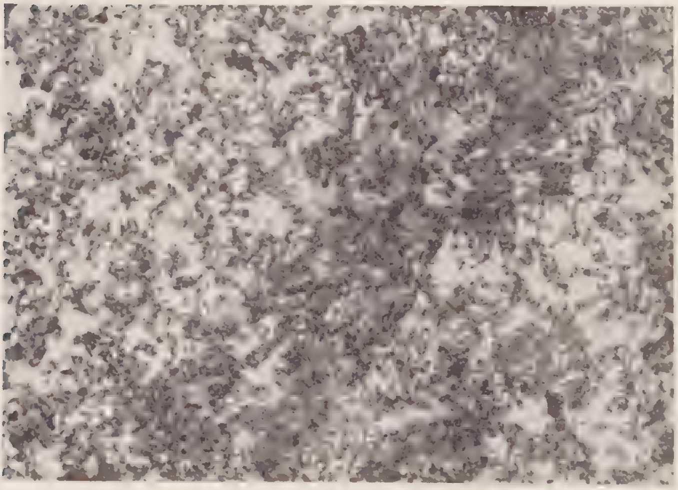
Chemical composition. The composition varies considerably, depending on the relative amounts of feldspar and dark minerals and on the kinds of dark minerals present. Table 4 illustrates that lime, iron, and magnesia are much more abundant than in granites. This abundance determines the larger content of hornblende and pyroxene, a feature characteristic of diorites.

PLATE 18.



A. ORBICULAR.

B. DIORITE,
NORMAL TYPE.



C. DIORITE, DARK TYPE.

TABLE 4

ANALYSES OF DIORITES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O^+	XyO[1]	Total	Sp. Gr.
I...	59.4	16.4	2.9	3.4	3.1	5.0	3.7	3.8	0.6	1.7	100.4	....
II...	52.1	18.0	4.1	6.3	5.0	8.6	3.0	1.6	1.0	0.7	100.5	2.89
III...	55.5	17.0	2.7	4.8	4.4	7.3	3.6	2.4	0.9	1.4	100.0	2.85

[1] XyO represents small amounts of other oxides.

I. Augite-hornblende-biotite diorite, Clayton Peak, Park City district, Utah.
II. Augite diorite, Stony Mountain, Ouray County, Colorado.
III. Average diorite, as represented by 14 analyses ranging in silica content from that of I to that of II (Rosenbusch-Osann).

Occurrence and uses. Diorites occur in many places as independent intrusive masses. They are also likely to be connected, on the one hand, as facies with granites, or, on the other hand, as facies with gabbros. They do not form such large masses as the granites or the gabbros, and, even where they occur as independent masses, they generally occur as stocks. They are widely distributed in all parts of the world.

Because of its dark color, diorite is not as extensively used for architectural purposes as granite, although, so far as strength, durability, and capacity for receiving a high polish are concerned, it would furnish excellent material in many places. Diorite is somewhat heavier than granite, its specific gravity ranging from 2.8 to 2.9; a cubic foot of diorite of specific gravity 2.85 weighs 178 pounds.

GABBRO

Composition. Gabbros are phanerites consisting chiefly of ferromagnesian minerals and plagioclase, the amount of ferromagnesian minerals equaling or exceeding that of the plagioclase. The megascopic distinction between gabbro and diorite is based on the relative proportions of ferromagnesian minerals and plagioclase: in gabbro the ferromagnesian minerals predominate, in diorite they are subordinate. The distinction between gabbro and diorite petrographically is based on the composition of the plagioclase: if the feldspar is a calcic plagioclase the rock is gabbro; if it is a sodic plagioclase, the rock is diorite; but this distinction, as was already pointed out, cannot be made in the field. Fortunately, the average gabbro carries more mafic minerals than the average diorite.

The ferromagnesian minerals are pyroxene, hornblende, and olivine, occurring either singly or together. Olivine if unaltered can be detected with the lens as yellowish or green grains. Biotite also occurs in some gabbros, but is far less common than in diorite. The pyroxene is generally black, appearing dark greenish when examined with the lens. Some is of the variety diallage, which has a pronounced apparent cleavage in one direction (actually, a parting), is gray green, and may be almost micaceous in appearance and have an almost semimetallic luster. A test with the knife point for cleavage shows at once that it is a nonmicaceous mineral. The feldspar in gabbros is a calcic plagioclase, generally labradorite. The twinning striae on cleavage surfaces can generally be seen, but the composition of the plagioclase cannot be determined megascopically. In many gabbros the plagioclase is dark colored, even almost black, so that it is not easily distinguishable from the ferromagnesian minerals. It is mostly in irregular grains as are the other minerals, but it may be tabular or booklike in form, and the sections as seen on the fracture surfaces of the rock are therefore elongate, with the twinning striae running parallel to the elongation. The feldspar may be fresh and glassy; its cleavage is then good, and the striae if visible are distinct. In many gabbros, however, the feldspar is waxy in appearance, of a glimmering luster to dull; its cleavage is poor or seems to be wanting, and striae cannot be seen. Feldspar of this kind has been more or less affected by alteration to other minerals, as described on page 35.

Gabbros are divided by petrographers into two groups, depending on the kind of pyroxene present. If the pyroxene is the monoclinic, calcium-bearing augite or diallage, the rock is called gabbro; if the pyroxene is the orthorhombic hypersthene, the rock is called *norite*. This distinction cannot be made megascopically, and identification would require that some of the pyroxene be extracted from the rock in question and be tested chemically or optically; hence norite is here included under gabbro. Norite is a rock of much economic interest, for at some localities, notably Sudbury, Ontario, it incloses immense bodies of nickel ore. A rarer variety of gabbro, consisting of plagioclase and olivine but containing no pyroxene, is called *troctolite*.

General properties. Most gabbros are dark: dark gray, greenish, black, or, rarely, reddish. Varieties in which diallage is the chief pyroxene and in which the grain is moderately coarse are lighter colored, being medium gray or greenish gray. The fabric is equigranular, but in many gabbros this fabric is modified by the tabular habit of the feldspars; porphyritic fabric is rare.

A banded or, more properly, a layered structure, produced by layers of unlike mineral composition superposed one upon the other and thus simulating stratification, is characteristic of many of the world's great gabbro masses—the Duluth lopolith, Minnesota; the Stillwater complex in Montana; and above all the Bushveld norite lopolith of South Africa, in which stratiform layering is extraordinarily well developed, rivaling that of most stratified sedimentary rocks.

Pegmatites are genetically associated with some gabbro masses. These pegmatites consist of the normal minerals of gabbro but are richer in plagioclase feldspar than the gabbro of the parent intrusive with which they are genetically associated.

Chemical composition. Gabbros contain much larger amounts of lime, iron, and magnesia and less silica and alkalies than any of the previously described rocks, as can be seen from Table 5.

TABLE 5

ANALYSES OF GABBRO

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O+	yO[1]	Total	Sp. Gr.
I...	53.5	15.4	1.6	8.8	5.7	8.5	2.8	2.2	1.2	1.2	100.1	2.90
II...	42.6	19.6	7.3	9.7	6.8	10.0	1.9	0.9	0.7	2.0	101.5	....
III...	47.4	18.5	3.3	7.3	6.8	11.0	2.4	0.7	1.3	1.2	99.9	2.92

[1] XyO represents small quantities and traces of other oxides.

I. Biotite-augite gabbro, Harzburg, Germany.
II. Biotite-hornblende gabbro, Val Scala, Veltlin, Switzerland.
III. Average gabbro, based on 20 analyses ranging from that of maximum silica content (I) to that of minimum silica (II) (Rosenbusch-Osann).

Occurrence. Gabbros are widely distributed. They occur as laccoliths, stocks, and batholiths and as dikes and sills. They also form the necks of volcanoes, as can be seen in deeply dissected volcanoes in regions of profound erosion. Gabbros occur also as the central portions of thick lava flows. This coarse-grained, gabbroic development in a lava is due to the low viscosity of such basic magma, which brings about its rapid and easy crystallization. Most geologists would not call such lava rocks gabbros, however, because to them "gabbro" is a plutonic rock; they would name them "coarse dolerites."

Alteration of gabbro. A gabbro mass that occurs in a folded belt of rocks is generally surrounded by a mantle of hornblende schist into which the central gabbro grades by transitional stages. This mantle was formed by shearing and other metamorphic processes active during the folding. The processes acted on the pyroxene of the gabbro, converting it into hornblende and producing the schistose structure. By forced flowage under shearing pressure a schistose or gneissic structure may be induced in the gabbro without its pyroxene changing to hornblende, and the resulting rock is a gabbro gneiss or gabbro schist; however, under these conditions the pyroxene is generally changed to hornblende. If olivine is present, it too is altered to amphibole. Garnet also appears as a new mineral resulting from the metamorphism. The change of the pyroxene to hornblende is usually, but not always, accompanied by the production of a gneissic or schistose structure; the transformed rock may retain its original massive granular character.

In another mode of alteration of gabbros the feldspar has been changed into a substance called saussurite, which was formerly thought to be a distinct mineral but has been shown by the microscope to be an exceedingly intimate mixture of albite and zoisite. Concurrently with the change of the feldspar to saussurite, the pyroxene was changed to a green amphibole. Other minerals also may have formed, but megascopically the waxy-looking saussurite and green hornblende predominate. This change may take place without the formation of a schistose structure; probably this alteration is due more to the chemical and less to the dynamic factors of metamorphism. A gabbro thus altered is called saussurite gabbro.

By weathering in a temperate humid climate, gabbros decay to clay soils deeply colored by hydrates of iron and mingled with fragments of still undecomposed minerals.

Iron and other ore deposits. Many of the large gabbro intrusive masses contain bodies of iron ore, consisting of magnetite and more or less ilmenite. Because of the titanium content of the ilmenite, these ores are called titaniferous iron ores. They are more or less mingled with the minerals of the gabbro itself, especially the pyroxene and olivine. The nature of the ore bodies and the manner in which they gradually grade into the normal gabbro surrounding them show that they are a facies of the gabbro, in which the magnetite and ilmenite, normally scattered through the gabbro in small grains, have become highly concentrated. Ore deposits of this kind occur at the borders of the intrusive bodies, scattered in masses through them, or even at their centers. Some of the ilmenitic magnetite masses are tabular and cut through the inclosing rocks, thus resembling dikes; in fact, they are considered by some to be dikes. Whether this interpretation is correct is at present a moot point, since it implies the existence of a magnetite magma, which would appear to have required, in order to make it molten, a temperature several hundred degrees higher than all evidence indicates that it could possibly have had. Such titaniferous magnetite bodies are known to occur in many places—the

Adirondacks, northern Minnesota, Canada, Norway, Sweden, and South Africa. For more than a hundred years it had been recognized that if titaniferous iron ores could be successfully smelted, many of these deposits would be of economic value. Rather unexpectedly, titanium became so valuable that in 1943 the large titaniferous iron deposits in the anorthosite of the Adirondacks were brought to commercial production for their titanium content; the iron in them is recovered as a by-product.

Sulfide ores have segregated in some gabbros in a similar manner. This is especially true of the iron sulfide pyrrhotite, which is nickel-bearing in some deposits and hence of value as a source of this useful metal. Some of these pyrrhotite deposits also contain valuable amounts of copper in the form of chalcopyrite. Such magmatic sulfide deposits in gabbros, or in rocks derived from them, have been worked in Norway, in Lancaster County, Pennsylvania, and possibly at Sudbury, Ontario, where, however, this interpretation is seriously questioned. As already mentioned, the nickeliferous pyrrhotite ore bodies are chiefly associated with the norite variety of gabbro.

Use of gabbro. Gabbros are well suited for constructional work and architecture, but they are not much used, probably in the main because of their dark color. In Sweden they have received considerable attention for monumental and other uses. In the United States they have been used as building stone in the Lake Superior region, as at Duluth, and quarries at Keeseville in the Adirondacks have been worked. Gabbros take a high polish, are sufficiently durable, and are much easier to work than granite.

Anorthosite. Anorthosite is composed wholly, or nearly so, of a plagioclase feldspar, generally labradorite. Pyroxene in small amount is sprinkled through it in grains, and many small masses of magnetite or ilmenite can be seen. This simple mineral composition makes anorthosite in the hand specimen, especially if it is rather fine grained and of light color, resemble both marble and quartzite, which also are rocks consisting of a single mineral. From marble it can easily be distinguished by its superior hardness, since feldspar cannot be scratched by the knife, whereas marble is easily cut or scratched; and from quartzite it can be distinguished by the cleavage of the plagioclase grains, which can usually be easily seen with a lens. Although these characters help to distinguish anorthosite, it can be identified with certainty only by determining that it contains plagioclase feldspar. This determination can be made in the field if the cleavage surfaces of the feldspars are large enough to permit one to see on them the characteristic twinning striae of plagioclase (see page 29). In the laboratory the feldspar can be identified by chemical and blowpipe tests, or more readily and precisely by optical tests.

Some anorthosites are white, but most are yellowish to brown, or more commonly some shade of gray—light gray, blue gray, smoky to dark gray, or almost black. The dark color is due to included pigment, perhaps minute plates of ilmenite. It is notable that where anorthosites have been sheared and granulated and have become gneissic the dark color tends to disappear and they become lighter colored.

The chemical composition of anorthosite is practically that of labradorite feldspar (analysis IV, page 33). Anorthosites that are composed of plagioclase more calcic than labradorite are potential ores of aluminum.

In mineral composition the anorthosites are related to the gabbros, for they contain the same kind of feldspar and many, as stated above, contain more or less pyroxene. If the pyroxene increases in amount, anorthosite grades into gabbro; on the other hand in some masses of gabbro facies poor in pyroxene occur and are thus transitional to anorthosite.

Anorthosite has two principal modes of occurrence: (1) as large independent intrusive masses—sovereign intrusive masses; and (2) as layers, from a fraction of an inch to many feet in thickness, as members of banded or layered gabbro lopoliths.

They are not common rocks, so far as the number of occurrences is concerned, but some of them form vast masses, notably in Canada and Norway. They occur in Canada in separate areas from the west coast of Newfoundland and the east coast of Labrador down through Quebec into eastern Ontario. An enormous area of anorthosite, drained by the Saguenay River, covers nearly 6000 square miles, and another, near Montreal, comprises 1000 square miles.

Another great mass is in the Adirondack Mountains of New York, where an area of 1200 square miles is underlain by anorthosite. Anorthosite occurs in considerable masses in Minnesota, on the north shore of Lake Superior, and in the Laramie Mountains, Wyoming.

Anorthosite is the most abundant representative of the group of igneous rocks called *monomineral* rocks, *i.e.*, rocks composed essentially of *one* mineral. It is regarded as having been formed by magmatic differentiation from a gabbro magma, whereby the labradorite crystals as they separated from the magma, being heavier than the magma, sank to the floor of the chamber and accumulated there. The labradorite crystals probably remelted at depth and so formed an anorthosite magma.

DOLERITE

Definition and minerals. Dolerite is a rock of gabbroic composition, which is intermediate in grain size between gabbro and basalt. "It is a basalt, whose components have become visible," wrote von Leonhard in 1823 in one of the first systematic treatises on rocks ever written, which perforce had to be based on megascopic methods.

In some dolerites the feldspar occurs as equant grains,* but in others it has a tabular shape, giving in cross sections laths of variant lengths; from microscopic studies the feldspar is known to be chiefly a calcic plagioclose. The ferromagnesian minerals are in dark grains, perhaps short columnar. Most grains show cleavage surfaces, but whether they are hornblende or pyroxene, or a mixture of both cannot be ascertained megascopically. Olivine also may occur, and if it is present, pyroxene and not hornblende is probably the chief ferromagnesian mineral. In addition, bronzy-looking flakes of biotite and metallic steel-like specks of magnetite or ilmenite can be seen with the aid of the hand lens.

Color. Because of their abundant ferromagnesian minerals, dolerites are dark—medium or dark gray or greenish to black. As in most rocks, the tone of color is best observed by viewing the rock at a little distance, so that the individual grains become indistinguishable and only their mass effect is seen.

General properties. Dolerites are equigranular; some are porphyritic, as described in the following section on the porphyries. The chemical composition of dolerites is like that of the gabbros. They are heavy, the specific gravity ranging from 2.9 to 3.3. Their jointing is characteristically columnar, the columns being on a grand scale in some places. Columnar structure is best developed in dikes, thick sills, and thick lava sheets—in tabular bodies that have cooled relatively slowly.

Occurrence. Dolerites do not occur in large stocks and batholiths as do the diorites and gabbros, though many gabbro masses grade into marginal facies consisting of dolerite. As intrusives the dolerites belong in the minor class, occurring as dikes, small laccoliths, and sills. They are exceedingly abundant as thick lava flows that have cooled slowly.

In the eastern United States well-known examples are the sills and flows of "trap" in the Triassic formations stretching from Nova Scotia to South Carolina. These dolerites form conspicuous topographic features, such as the ridges in Connecticut and the Palisades on the west

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* A grain is said to be equant if its three dimensions are roughly equal.

side of the Hudson River across from New York City. Similar great sills of dolerite occur in the Lake Superior region and in the great lava flows of the western United States.

Dolerites are also common rocks in dikes and sills in Great Britain, in northern Scotland, and in Ireland. Dolerites are prevalent, in fact, in all parts of the world.

Relation to other rocks. From what has been said it is easy to see that the dolerites are a class of rocks based largely on practical exigency. They form a transitional group, based on grain size, between gabbros and aphanitic basalts. Hence there is no sharp boundary between finely phanerocrystalline gabbro and dolerite, and between dolerite and basalt. Dolerites are logically called by the French *microgabbros.**

Dolerites, by severe metamorphism, have become converted into hornblende rocks, generally into hornblende schists. Under less severe conditions they have been altered into "greenstones." These alterations are quite similar to those described under gabbro. Dolerites by weathering become brownish and discolored and ultimately yield brown ferruginous soils.

Uses. Dolerites are too dark and somber for general use in fine architectural or interior work, except for monumental purposes. The "trap" of the Eastern States has been much used in rough masonry, and where good natural joint faces can be used for wall surfaces, the brown weathering color gives pleasing effects. The toughness of trap endows it with a strong resistance to abrasion and has caused it to be extensively used for block paving and as crushed stone for road making and concrete mixture.

PERIDOTITE—PYROXENITE—HORNBLENDITE

Composition. In this section are discussed all equigranular phanerites composed solely of ferromagnesian minerals and those in which the associated feldspar is so small in amount as to be negligible. They are ultramafic rocks, generally holomafic. The chief minerals are olivine, pyroxene (both clinopyroxenes and orthopyroxenes), and hornblende. The minerals occur singly or in various mixtures, and on the basis of the minerals present the rocks are subdivided into types, the more prominent of which are as follows:

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* In America intrusive dolerite of diabasic fabric is commonly called *diabase* under the belief that this fabric occurs only in intrusive masses. However, many lavas have diabase fabric, in fact, many flows from Kilauea have diabasic groundmasses, in flows only 12 feet thick.

Olivine (predominant) and pyroxenes	Pyroxene peridotite
Olivine (predominant) and hornblende	Hornblende peridotite
Olivine alone	Dunite
Pyroxenes alone or predominant	Pyroxenite
Hornblende alone or predominant	Hornblendite

The first three, in which olivine predominates, are comprised under the group name of peridotite, from *peridot*, French for olivine. All the different peridotites, although occurring in places as individual intrusive masses of homogeneous composition, also occur associated together within a single intrusive mass, with transitional facies grading into one another; it is difficult, and generally impossible, to distinguish them megascopically.

Besides the minerals mentioned, brown biotite may occur in peridotite, giving rise to the variety called *biotite peridotite*. The chief accessory mineral of the peridotites is chromite in small sharply defined octahedrons.

Texture. The fabric is equigranular, but its appearance is modified somewhat by the minerals present and their arrangement. If pyroxene or hornblende is dominant, the grain is likely to be coarse and large cleavage surfaces are visible. Dunite, however, is generally not coarse grained but has a sugar-granular texture like many aplites. Porphyritic fabric is rare or wanting in all the holomafic rocks. Characteristically, the pyroxenes and hornblendes as seen on their cleavage surfaces are spotted with inclosed grains of olivine. Such spotting of the cleavage surfaces of one mineral by enclosed crystals of another mineral that are scattered in random orientation throughout its host is called *luster mottling*, or *poikilitic texture*. It is strikingly developed in some peridotites, notably in bronzite peridotite. The enclosed crystals of olivine seen in the luster-mottled patches of bronzite are, of course, older than the bronzite, their host mineral.

Chemical composition. The chemical composition of the holomafic rocks varies according to the mineral composition, but the typical features are low silica, high magnesia, considerable iron, and virtual absence of alkalies and alumina. Because of the high proportion of the bases magnesia and iron, these rocks are often called *ultra-basic*.

Color. Peridotites range from dull green to black. Dunites, being monomineral rocks composed of olivine, are much lighter in color. They are light green and yellowish green and accordingly are exceptions to the rule that ferromagnesian rocks are dark colored.

Occurrence. Relation to gabbro. Peridotites and allied rocks occur independently as dikes, sills, laccoliths, and stocks. In a few places

TABLE 6

ANALYSES OF HOLOMAFIC ROCKS

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	+XyO[1]	Total	Sp. Gr.
I...	40.1	7.8	7.3	8.6	23.7	6.5	1.2	0.5	4.0	0.6	100.3	..
II...	43.9	1.6	8.9	2.6	27.3	6.3	0.5	...	8.7	0.3	100.1	..
III...	53.2	1.9	1.4	7.9	20.8	13.1	0.1	0.1	0.9	0.9	100.3	..
IV...	46.4	10.8	5.9	5.6	22.2	3.7	0.3	1.2	3.8	...	99.9	..
V...	42.8	...	...	9.4	47.4	...	....	...	0.6	1.4	100.2	3.30

[1] XyO represents small amounts of other oxides.

I. Peridotite, Devonshire, England.
II. Peridotite, Baltimore County, Maryland.
III. Pyroxenite, Oakwood, Maryland.
IV. Hornblendite, Valbonne, Pyrenees.
V. Dunite, Mount Dun, New Zealand.

they form masses of batholithic dimensions: in New Caledonia there is a mass of 1400 square miles, and still larger masses are reported to occur in Cuba. They occur also as stratiform layers in gabbro lopoliths, notably the Duluth and Bushveld lopoliths. In this mode of occurrence the peridotite layers are clearly products of differentiation of the gabbro magma.

Dunites occur in masses intrusive into the gneisses of western North and South Carolina and Georgia. Associated with them are smaller amounts of other peridotites and pyroxenite. Within the last few years these dunite masses have become of commercial interest as sources of refractory brick. Dunite occurs also in considerable masses in the Ural Mountains of Russia, where the largest mass, that of Nizhny Tagil, has an area of 17 square miles. The type locality of dunite is Mount Dun, New Zealand, so named from the characteristic brown or dun color to which its dunite weathers.

Pyroxenite is less common than peridotite, but in places it occurs in immense volume, notably the variety consisting wholly of bronzite (bronzitite) which forms a layer 6000 feet thick in the Bushveld lopolith of South Africa. Hornblendite is comparatively rare.

Alteration. Serpentine. Peridotites are altered so commonly, that fresh, unchanged rock is rare. In the most prevalent form of alteration the olivine and other magnesian silicates have been changed to serpentine. Such serpentinized peridotite is called *serpentine*, the name serpentine being thus used both as the name of a mineral and of a rock. Recently the tendency is to call the rock *serpentinite*.

Other magnesian minerals, such as talc, are also formed by the alteration of peridotites and pyroxenites, but the change to serpentine is by far the most widespread. All stages of transition from unaltered peridotite and pyroxenite to rock composed wholly of serpentine occur, and most serpentinites the world over have been formed by the alteration of peridotites and pyroxenites.

Serpentinites resist weathering but ultimately break down into brown ferruginous soils, which, because of their lack of potash, do not favor the growth of vegetation and are therefore barren.

Source of valuable minerals. The magmas from which the peridotites formed carry small amounts of chromium which crystallize out in the form of chromite ($FeCr_2O_4$), one of the spinel group. Chromite is common in dunite and forms small, black, brilliantly lustrous octahedrons. In some peridotites it has become sufficiently concentrated to form valuable ore bodies, supplying the chromium used in the industries. All of the world's chromium comes from chromite bodies inclosed in peridotite and pyroxenite, but all peridotite and pyroxenites do not contain chromite ore. The olivine of peridotites contains a minute amount of nickel, and when this olivine weathers the nickel may under the requisite conditions become concentrated in the form of nickel silicate, in amounts sufficient to form ore bodies of value as a source of nickel, as in Douglas County, Oregon, and above all on the island of New Caledonia, the world's second principal source of nickel.

The peridotites, and to a minor extent their allies the pyroxenites, are also in some localities a source of platinum, which occurs in them as the native metal. By the decay of the host rock the platinum is liberated and is washed down into streams and, like gold, becomes concentrated in the stream gravels. The most notable examples are the platinum-bearing placers of the streams that head in areas of peridotite and pyroxenite in the Ural Mountains of Russia.

The diamonds of South Africa are obtained from highly serpentinized rock, which constitutes the necks of old volcanoes. This altered rock, formerly thought to have been originally a peridotite, is now known as *kimberlite*, a rarer and more specialized type of igneous rock than peridotite. The carbon forming the diamonds was at first believed to have been derived from the carbonaceous shales through which the kimberlite magma passed on its way up to the Earth's surface; but it is now thought that the carbon was an original constituent of the kimberlite magma and that the diamonds crystallized from this magma just as did the other minerals. The reason for this change of belief is that during the downward progress of mining the kimberlite pipes for diamonds, the diamond content persisted far below the level of the carbonaceous shales.

PORPHYRITIC ROCKS*

Definition. Porphyritic rocks are characterized by the presence of phenocrysts scattered through a groundmass. The phenocrysts range from almost none to practically 100 per cent of the volume of the rocks. In most of these rocks the groundmass is finely phanerocrystalline, microcrystalline, or aphanitic. Those having a phanerocrystalline groundmass and those having more than 50 per cent of phenocrysts regardless of the grain size of groundmass, whether aphanitic or phanerocrystalline, comprise a group called *phanerite porphyries*. The remaining porphyritic rocks—actually an enormous group—are called *aphanite porphyries* (aphanophyres). Thus two groups are obtained, one in which phanerocrystalline matter predominates and a second in which aphanitic matter predominates.

The naming of the phanerite porphyries is relatively simple. It is done by taking the name of the phanerite to which a given porphyritic rock corresponds in composition and adding the term porphyry: thus the name *granite porphyry* implies that the rock so designated has the composition of granite, has a porphyritic fabric, and has a phanerocrystalline (generally finely phanerocrystalline), or microcrystalline groundmass.

Porphyritic rocks of this group are mainly of silicic composition. The reason for this limitation in composition is that subsilicic magmas have a low viscosity, and hence the minerals crystallize so readily from them that they do not, except rarely, form porphyritic rocks, whereas the silicic magmas, cooling under the same conditions, almost invariably form porphyritic rocks. Granite porphyry is consequently common, but gabbro porphyry and peridotite porphyry are practically nonexistent. Accordingly, the rocks to be described are granite porphyry, syenite porphyry, and diorite porphyry. They have so many points in common that they can be described as a group.

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* The phrase "porphyritic rock" is used instead of the apparent synonym "porphyry," because "porphyry" means different things to different people; it may mean (1) an intrusive porphyry, regardless of the grain size of the groundmass, as in the system of the distinguished petrographer Rosenbusch; (2) an intrusive rock regardless of whether it contains phenocrysts or not—rhyolite porphyry is a name for rocks of rhyolitic composition and fabric if intrusive *even though not porphyritic;* or (3) an abundantly porphyritic rock—rhyolite porphyry in this usage means a rhyolite containing between 25 and 50 per cent of phenocrysts; recently the last definition has been redefined to mean between 10 and 50 per cent of phenocrysts.

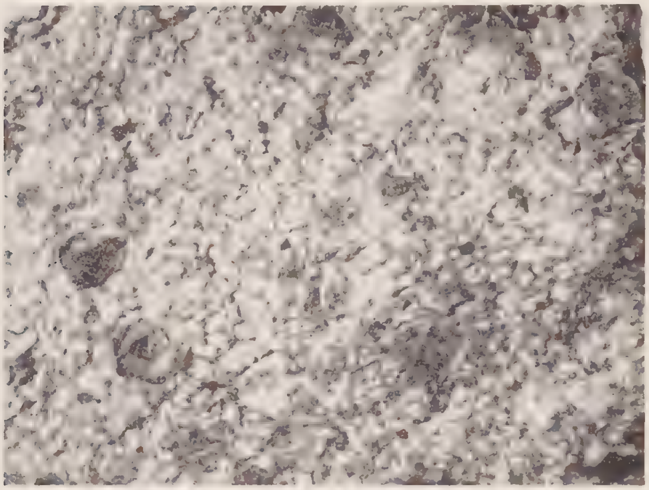
Granite porphyry. Granite porphyry is characterized by containing phenocrysts of feldspar and quartz in a groundmass of the same minerals. In many granite porphyries the groundmass is finely phanerocrystalline or verges on aphanitic, *i.e.*, it is microcrystalline. In some granite porphyries the feldspar and quartz phenocrysts are so numerous that they give a distinct granite-like aspect to the rock, even though the groundmass is aphanitic. Porphyries that contain more than about 50 per cent of phenocrysts are of granite-like aspect, and those containing more than 75 per cent have so little groundmass that it is not discernible by the unaided eye, and consequently these porphyries cannot be distinguished megascopically from granite. In Plate 19 is shown a granite porphyry carrying large but sparse phenocrysts of quartz and abundant smaller phenocrysts of feldspar.

Granite porphyry may consist wholly—or almost wholly—of feldspar and quartz, but biotite and hornblende also may be present in minor amount. The biotite and hornblende occur separately or together, though hornblende without biotite is rare. The biotite and hornblende occur as distinct phenocrysts generally smaller than those of feldspar and as a rule smaller than those of quartz; and they may also occur in minor quantities in the groundmass, where only the tiny flakes of biotite are easy to detect. Orthoclase is the predominant feldspar, and some sodic plagioclase is almost invariably present. As the proportion of this associated plagioclase increases, granite porphyry grades by imperceptible steps into rocks termed in more refined classification quartz monzonite porphyry and granodiorite porphyry. These porphyries are intermediate in composition between granite porphyry as precisely defined and quartz diorite porphyry.

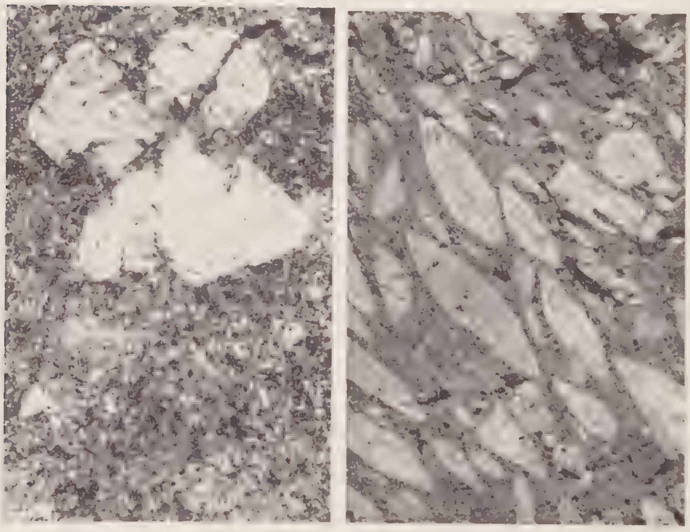
Porphyries of these intermediate kinds have in recent decades become of great economic interest, inasmuch as in a mineralized state they constitute the ore at many of the great disseminated copper deposits that have revolutionized the copper industry during the twentieth century, as, for instance, the monzonite porphyry at Bingham, Utah; at Santa Rita, New Mexico; and at other great copper-producing districts.

Syenite porphyry. Syenite porphyry consists essentially of phenocrysts of orthoclase in a phanerocrystalline or aphanitic groundmass made up of grains of feldspar with little or no quartz. If the groundmass is aphanitic, the amount of phenocrysts must exceed 50 per cent, giving the rock a distinctly syenitic aspect. The ferromagnesian minerals—biotite, hornblende, and pyroxene—though they may be absent or practically so, are generally present, either as phenocrysts or in the groundmass, or in both forms. They occur separately or together,

PLATE 19.



A. GRANITE PORPHYRY, MONTANA.



B. SYENITE PORPHYRY, GERMANY.

C. SYENITE PORPHYRY, NORWAY.

but the association of all three or of biotite with pyroxene is not as common as that of biotite with hornblende.

The syenite porphyry just defined corresponds in composition to the commoner of the two kinds of syenite described on page 181 and following. It is not a common rock. Nepheline syenite porphyry is also known but is very rare.

Diorite porphyry. Diorite porphyry is composed of phenocrysts of plagioclase and of biotite, hornblende, or pyroxene, either separately or together, in a finely phanerocrystalline or microcrystalline groundmass, which is chiefly plagioclase. If the groundmass is not phanerocrystalline, the diorite character must be clearly shown by the abundance of the phenocrysts, which should exceed 50 per cent of the bulk of the rock.

If quartz also occurs among the phenocrysts, the rock is called *quartz diorite porphyry.* Porphyry of this kind is likely to resemble granite porphyry and the varieties intermediate between granite porphyry and quartz diorite porphyry, namely, quartz monzonite porphyry and granodiorite porphyry. The old term *tonalite porphyry* has recently been revived as a synonym for quartz diorite porphyry.

Phenocrysts of porphyries. As the phenocrysts of porphyries have crystallized freely in fluid magmas, they are generally well-shaped crystals; in technical parlance, they are *euhedral* crystals, such as are described in Chapter IV. A few words as to their crystal habits may be added there. Quartz phenocrysts tend to have the bipyramidal form shown in Fig. 41, somewhat modified, however, so that they approximate ellipsoidal or spherical shapes. The crystals may be as large as 1/2 inch in diameter, but most are smaller, being about the size of coarse shot or peas. The feldspars tend to have the forms shown in Figs. 3 to 5; many are twinned, as in Fig. 6; they are white, pink to red, or yellowish and gray. If feldspars of two colors are present and one of them is reddish, it is probably orthoclase, and the other is albite or a sodic plagioclase. They may form large phenocrysts, an inch or even more in length. Many of the model feldspars seen in mineral cabinets are phenocrysts obtained from porphyries. Hornblende occurs in dark-greenish or black prisms, usually elongate; some are slender and needle-like and have brilliant cleavage surfaces if fresh; the terminal faces are poor or wanting. The hornblende may have weathered out, and only a rusty mass remains. Pyroxene also is dark green to black; it is in short, stout prisms, and its cleavage and crystal faces generally lack the brilliant luster of hornblende. The method of distinguishing between hornblende and pyroxene has already been explained. Both hornblende and pyroxene tend to be small compared with feldspar. Biotite, as phenocrysts, is in six-sided tablets having a fine basal cleavage; it is black to bronze brown. In these rocks its crystals are mostly small.

General properties. The chemical composition of the porphyries is like that of the corresponding kinds of equigranular phanerites. Their specific gravity, modes of alteration, and conversion into soil are similar. The jointing depends largely on the mode of occurrence; it is generally platy or small cuboidal, or it forms small parallelopipedons with acute angles in the silicic porphyries of dikes and sills, and larger blocks in the greater intrusive masses. Doleritic porphyries tend to have columnar jointing. It must be emphasized that columnar structure is not restricted to dolerites, however, but occurs in rocks of all compositions, including rhyolite.

Occurrence. The phanerite porphyries commonly occur as intrusive masses of relatively small volume; as dikes, sills, small laccoliths, and volcanic necks. They occur also as the border facies of many stocks and batholiths of granite and allied rocks; here they represent a contact modification produced by rapid chilling. They occur also in lava flows, especially if these are thick and massive, but in this mode of occurrence, because of the necessarily rapid cooling, the felsite porphyries and basalt porphyries described in later paragraphs are more likely to occur.

It is impracticable to give a list of localities for these rocks, because they are far too common. They are found wherever erosion has exposed the older crystalline rocks and where igneous activity has taken place. Where there are large stocks and batholiths, microphanerite porphyries are especially likely to occur. Some cut the stocks and batholiths as dikes, some extend from them as offshoots, and others occur as dikes and sills in the surrounding rocks.

Notable occurrences of microphanerite porphyries are in the laccoliths of the Rocky Mountain region, in Colorado, Utah, Wyoming, and Montana. The laccoliths are generally composed of granite porphyry or syenite porphyry, and some of them are a mile thick and several miles in diameter, though many are smaller. In these States the porphyries also form large sills, hundreds of feet thick.

APHANITIC IGNEOUS ROCKS

In the preceding groups of igneous rocks all or most of the component grains of the rocks can be determined megascopically and the mineral composition can therefore be ascertained. The group of aphanitic rocks comprises those in which the grain size of all or most of the rock is so minute that it cannot be ascertained. No definite line can be drawn between the two groups. In the nature of things there is bound to be a twilight zone, and whether some rocks in the transition zone between phanerocrystalline and aphanitic belong in one category or the other, is a matter of opinion, dependent on the experience of the observer and his power of observation.

The aphanites, as already explained in the section on the classification of igneous rocks, fall into two groups on the basis of color, *i.e.*, into the light-colored felsites and the dark basalts. Each of these groups has a nonporphyritic and a porphyritic subdivision. The felsite group will be described first.

FELSOPHYRE (FELSITE PORPHYRY) AND FELSITE

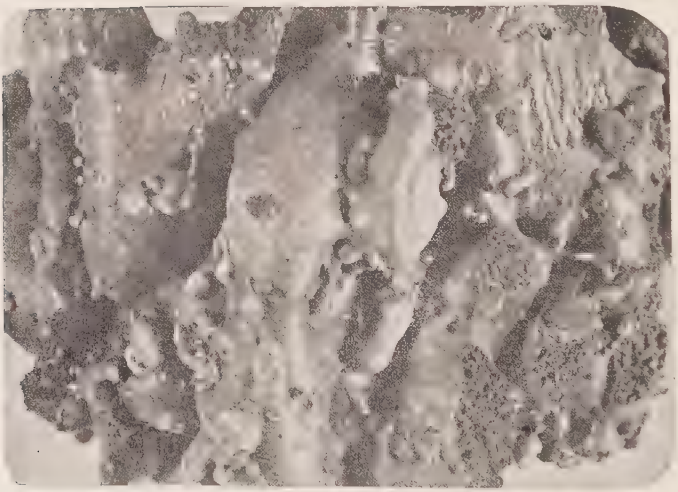
Felsite porphyries, or felsophyres, include all extremely fine-grained light-colored porphyritic igneous rocks of stony texture. If devoid of phenocrysts, they are termed felsites. However, the porphyritic rocks, the felsophyres, vastly predominate, so that it seems more logical to use "felsophyre" rather than felsite as the group name. They are normally light in color: white, which is not common, light to medium gray, light pink or red to dark red, pale yellow or brown, purple, or light green; in short, all colors except dark gray, dark green, and black. With the lens they are generally seen to consist of minute mineral grains, too small to identify, and the texture is then very finely granular. The grains may be too minute to be perceived, and consequently the rock has a hornlike or flinty aspect, seeming to be a homogeneous substance, and is likely to have a smooth, conchoidal fracture. Again, especially in lavas, the texture is more or less porous and the fracture surface of the rock is rough and hackly, with a harsh feeling. Vesicular structure, prevalent in basalts and in glassy rocks, and illustrated in Plate 8, is not common in the felsites. Many show fluidal banding and streaking, in places curved or curled, because of flowage; these flow structures are clearly brought out on weathered surfaces, as illustrated in Plate 20.

Aphanites are divided into felsophyres and basalts on the basis of color, and not on mineral composition, since their composition cannot be determined megascopically. Nevertheless, the felsites as thus classed on the basis of light color are in general silicic rocks, and they represent in aphanitic form, chiefly as lavas, the magmas that under different physical conditions of cooling and solidification would have produced granites and syenites. The basalts correspond chiefly to gabbros and dolerites, as already explained in the classification.

If aphanites are medium gray or drab in color, they are with difficulty assigned either to the felsites or the basalts. If a felsite is

PLATE 20.

A. FLOW STRUCTURE IN GLASSY LAVA.



B. SLAGGY STRUCTURE OF BASALTIC LAVA.

closely examined with a good lens, many tiny chips and flakes, only partly formed and yet in the main a part of the mass, can be seen to lie upon its surface, their thin edges separated from it by a flat underlying crack. The thin edges of these flakes are much lighter in color than the rock upon which they lie and are translucent to light; indeed, they may appear to be practically white, even though the rock is a dark gray. This translucence is peculiar to the aphanites that are composed chiefly of feldspar and therefore are to be classed as felsites; it is not shown, or shown only in a very small degree, by the basalts. The reason for this phenomenon is that the feldspars are transparent or translucent minerals, even though colored by a pigment, and consequently a rock composed mostly of feldspars is megascopically translucent on thin edges, whereas one composed chiefly of ferromagnesian minerals is not, because the ferromagnesian minerals are megascopically either opaque or practically so. The same translucent effect can be observed on the very thin edge of a flat chip broken from the rock. If the individual grains are large enough to be seen with the lens and are transparent or translucent with light colors, they can be assumed in general to be mainly feldspar, as the grains of ferromagnesian minerals are mostly dark and dull to opaque. Olivine, however, being yellow green, is an exception and must not be confused with feldspar. It will be noted also that the mineral grains, or many of them, appear to be much lighter in color under a good lens than the general effect the rock makes on the eye.

These tests often serve in helping to decide whether a rock is a felsite or a basalt but are not exact, for the distinction between these rocks, although based on color, is fundamentally due to the difference in mineral composition.

It is assumed in what has been said regarding these rocks that one is dealing with fairly fresh, unaltered material, not rocks that have been modified by weathering.

Varieties of felsite. Varieties of felsite are recognizable megascopically on the basis of color and texture. By the use of the microscope on thin sections, however, petrographers are able to determine the mineral composition of the minute grains and to classify the felsites on the basis of their differing mineral compositions, just as they have done with the coarse-grained rocks, the grains of which can be seen and determined by the unaided eye.

Felsite porphyry, or felsophyre. For the purposes of megascopic classification the upper limit of the amount of phenocrysts in the felsophyres is placed at 50 per cent of the volume of the rock. Phenocrysts range accordingly from few or none, as in felsites, to very many in others. The phenocrysts are chiefly sialites—quartz and feldspar—with which one or more mafic minerals—biotite, hornblende, or pyroxene—may occur.

In general, phenocrysts of more than one kind are likely to be present. Quartz and feldspar may occur along with phenocrysts of the mafic minerals, either singly or in the various combinations possible. However, biotite is the mafite most likely to occur along with quartz phenocrysts. Hornblende phenocrysts in association with quartz are less common, and pyroxene phenocrysts with quartz are rare. Feldspar and hornblende phenocrysts are common associates, as are pyroxene and plagioclase. Feldspars, including both orthoclase and plagioclase, are the most prevalent phenocrysts.

Felsite porphyry is the name given to these rocks, a name that, as already indicated, can be shortened to *felsophyre* (from *fels*, from felsite, and *phyre*, shortened form of porphyry), or to emphasize the light color the name *leucophyre* (Greek *leucos*, white) is given to them.* If no mineral qualifier is used with felsophyre in designating a rock, it implies that the rock has phenocrysts of feldspar, since the presence of feldspar phenocrysts is the general rule.

If in addition to feldspars the rock contains phenocrysts of quartz, biotite, hornblende, etc., we have the following terms:

Quartz felsite porphyry, or *quartz leucophyre*.
Hornblende felsite porphyry, or *hornblende leucophyre*.
Biotite felsite porphyry, or *biotite leucophyre*.
Augite felsite porphyry, or *augite leucophyre*.
Hornblende-feldspar leucophyre.
Quartz-feldspar leucophyre.

These examples suffice to show how mineral qualifiers can be used to indicate what phenocrysts are present in these felsites, and how these qualifiers can be combined, if desired, to express considerable descriptive detail. Other examples readily suggest themselves. An outstanding virtue of these names is that they frankly proclaim that they are megascopic designations, given without benefit of the microscope.

General properties. The chemical composition of felsophyres is variable, depending, as in the varieties enumerated above, on the prevailing minerals. Some analyses correspond to those of granite already given; the corresponding felsophyres contain free quartz. Other analyses are like those of syenites, and the corresponding rocks

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* As suggested in Cross, Iddings, Pirsson, and Washington, *Quantitative Classification of Igneous Rocks*, University of Chicago Press, 1903, p. 184.

contain little or no quartz. If lime is small in amount, silica is below 60 per cent. and alkalies are high, the rock must be mostly composed of alkali feldspar and nepheline is probably present.

The specific gravity ranges from 2.4 to 2.65 and is generally lower than that of granite and syenite. The jointing is platy, or in small blocks. Columnar structure also occurs, but is not so common as in basalts. In normal weathering to soil these rocks become discolored to brownish or reddish tones. The ferromagnesian mineral generally disappears, leaving a rusty spot or cavity; the rock crumbles into debris, at first largely through mechanical disintegration. Finally, the feldspars alter to kaolinite or other clay minerals, as described under granite, and the change to soil in a temperate region is complete. In tropical regions, under the requisite climatic and topographic conditions, the changes go one stage further. The kaolinite is desilicified, the liberated silica is carried away by the ground-water circulation, and the final product is a surface layer of more or less ferruginous laterite.

Where hydrothermal action has taken place, as in the wall rocks adjacent to ore veins, the felsites have commonly been altered to whitish rocks containing abundant sericite, the finely scaly form of muscovite.

Occurrence. Felsites occur as dikes and sills, and in places as the chilled border facies of other intrusive masses whose interior portions consists of granite, granite porphyry, syenite, or some such rock. Such felsitic borders indicate that the magma was injected into cold rocks and was drastically chilled, so that the minerals were unable to grow to more than minute size.

Felsites and felsophyres are much more common as lava flows. Some of them cover large areas, hundreds and even thousands of square miles in extent. In these occurrences and especially in volcanic cones the lavas are generally intercalated between beds of tuff and breccia.

It is impracticable to give a list of occurrences, but it may be mentioned that felsophyres as intrusives occur extensively in eastern North America, among the older rocks in the Atlantic border States; as extrusives they occur in Maine, in the White Mountains, in Pennsylvania, and southward. Much more voluminous is their occurrence as lavas throughout the whole of the Cordilleran tract in North and South America where they make up the bulk of many of the ranges and occupy large areas. Here all the many varieties are found, notably rhyolite in Yellowstone Park, in Colorado, and elsewhere, and andesite in the lofty volcanoes of the Cascade Range, Mexico, Central America, and the Andes of South America; the rare variety phonolite occurs at Cripple Creek, Colorado, and in the Black Hills of South Dakota.

Felsophyres are abundant in other parts of the world. In fact, wherever extensive volcanic outbreaks have taken place, felsitic lavas occur.

Felsites and felsite porphyries occur as narrow dikes that cut the larger stocks and batholiths of phanerites, such as granite and syenite, and extend out into the rocks surrounding these intrusive masses. These felsitic dikes are generally complementary to dark dikes, called lamprophyres. They have received special names: *bostonite* (allied to trachyte), *tinguaite* (allied to phonolite), and many others. In this connection the paragraphs on complementary rocks, on aplite, and on lamprophyres should be reread.

BASALT

Basalts comprise dark aphanitic igneous rocks, the fabric of which is so fine grained that the constituent grains either are not perceptible by the unaided eye or with the lens, or, if perceptible, are too small to be identifiable. They have a stony, not a glassy, texture. They range from grayish black or dark stone color through greenish or purplish black to pure black. As a rule basalts are not translucent on the edges of thin flakes, differing thus from felsites. If not cellular and if very fine grained, they have a uniform dull, soft, almost velvety appearance and do not show the horny or flinty luster characteristic of most felsites.

In thin section basalts are seen to be composed of minute crystal grains of plagioclase feldspar (generally labradorite), pyroxene, and iron ore; some contain more or less olivine, and some contain biotite or hornblende. Nepheline or leucite may accompany the feldspar, or proxy for it, giving rise to varieties that have received special names. These varieties, although interesting from the standpoint of theoretical petrography, are comparatively rare.

Being composed of the same minerals, basalts represent in aphanitic form and generally as lavas those magmas which under different physical conditions would have solidified as gabbros, peridotites, dolerites, and (in part) diorites. Dolerites by definition are rocks intermediate in grain size between basalts and gabbros, as previously mentioned in the description of dolerite, and examples are common where in the same rock mass the gradation from basalt to dolerite or from dolerite to gabbro can be seen.

Porphyritic basalts are abundant and widely distributed, but porphyritic fabric is far less common in this group than in felsites. This fact is undoubtedly the result of the low viscosity of basaltic magma and the consequently easy crystallization of the minerals. One exception, however, is the presence of olivine, which is likely to occur scattered through the basalt in transparent yellowish- or bottle-green porphyritic grains, averaging the size of moderately coarse shot. Olivine is so common (indeed at one time a rock was not considered to be a basalt unless it contained olivine) and so rarely produces a striking porphyritic effect, that the custom is to ignore its porphyritic effect, and to call basalts that contain olivine phenocrysts *olivine basalt* rather than olivine basalt porphyry.

The chief minerals that occur as phenocrysts are olivine, as already mentioned, pyroxene, and plagioclase; hornblende and biotite are far less common. The plagioclase is commonly labradorite; it is in long tabular crystals, either singly or in twinned groups. The pyroxene is chiefly augite; it is shining or dull black and is in short, thick prisms, as illustrated under pyroxene. The hornblende also is black and is brilliantly lustrous on its excellent cleavage surfaces. Biotite is in six-sided tablets.

These rocks are named in accordance with the prevailing phenocryst, for example, *augite basalt porphyry*. Instead of the name basalt porphyry, *melaphyre* ("black porphyry") can be used as an appropriate megascopic name. Accordingly we have *feldspar melaphyre, augite melaphyre, biotite melaphyre,* etc.

General properties of basalt. The chemical composition of basalt varies with its mineral composition; it is like that of gabbro previously given, as shown by the following average analysis of basalt (198 analyses averaged by R. A. Daly).

SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
49.1	15.7	5.4	6.4	6.2	9.0	3.1	1.5	1.6	2.0	100.0

[1] XyO represents small amounts of other oxides.

The specific gravity is high—about 3.0 (2.9 to 3.1). The jointing is platy or columnar; indeed, the finest examples of columnar structure are seen in basalt. Many notable examples are to be seen in all parts of the world, the Giant's Causeway on the north coast of Ireland being one of the best known. An impressive example is shown in Plate 11. Some basalts on weathering have developed a remarkable "spheroidal" structure, consisting of rudely spherical, onion-like, exfoliating masses.

Varieties. Among the fine-grained nonporphyritic basalts there is little diversity of appearance, save that based on change from compact to cellular, or vesicular, structure. Vesicular structure is particularly common in lava flows, especially in their upper and their bottom portions; it is illustrated on Plate 8. Amygdaloidal structure, illustrated on the same plate, is especially prevalent in these basalts. The minerals that fill the cavities in basalt are commonly quartz, calcite, chlorite, and zeolites. Analcite, natrolite, stilbite, and heulandite are the most prevalent of the zeolites. Such rocks are called *amygdaloidal basalt.*

In some places, particularly in western America, basalts containing visible grains of quartz occur as lavas. One of the most noted is the basalt flow from Cinder Cone, in Lassen Volcanic National Park in northern California, which is filled with angular pieces of quartz of various sizes. As many of these basalts correspond in composition to gabbros and dolerites, the presence of quartz in them is anomalous, since magmas so low in silica, as can be seen by referring to the analyses of gabbro, would not be expected to develop free quartz on solidifying. The evidence points to the conclusion that the quartz was derived from fragments of quartz rock brought up from the depths, which were torn loose, distintegrated, and distributed through the magma. These quartz-bearing basalts have been called *quartz-basalts.*

A variety of basalt characterized by abundant phenocrysts of labradorite is called *labradorite porphyry;* another variety containing abundant phenocrysts of augite is called *augitophyre,* or more adequately, augite melaphyre. (See Plate 21.)

The name *trap* is often used as a field term to designate the rocks here called basalts and dolerites. So used, it means any dark-colored, heavy, igneous rock of undetermined mineral composition. Thus the basalts and dolerites in the Triassic beds along the Atlantic Coast have been called "Triassic traps," and the great mass of basaltic lavas of western India are known as the "Deccan traps."

Lamprophyres. The ferromagnesian complementary rocks that occur as dikes and sills in or around stocks and batholiths of granite and other major intrusive masses are often called traps, but are technically *lamprophyres.* Their relations and origin have been discussed in the preceding part of this work. Generally the lamprophyres contain phenocrysts of ferromagnesian minerals—olivine, augite, hornblende, and biotite—either singly or together. The phenocrysts are embedded in a groundmass commonly aphanitic and dark colored. Many different lamprophyres have been named by petrographers

PLATE 21.

A. Labradorite Melaphyre.

B. Augite Melaphyre.

VARIETIES OF PORPHYRITIC BASALT.

according to the minerals present, as shown by the microscope, but the types thus recognized cannot as a rule be distinguished megascopically. For field work they can be treated as melaphyres as described above, and called *augite melaphyre*, *biotite melaphyre*, etc. Biotite melaphyre frequently occurs as dikes associated with granite, and is called *mica trap* (technically, *minette*). A hornblende melaphyre is associated with many syenites and nepheline syenites; it occurs in New England, where it was first discovered, and has since been found the world over; it is called *camptonite*.

Olivine segregations or dunite xenoliths. Some basalts, in addition to their phenocrysts of olivine, inclose yellowish or green masses made up of grains of olivine. Grains of pyroxene and spinel also may be present in these enclosures. The enclosures range in size from a pea to bodies as large as one's fist, or even larger. Some are round, but many are distinctly angular in shape. The origin of these enclosures is problematic. Some authorities hold that they are segregations formed by the assembling of earlier-formed crystals of olivine that separated from the basalt magma, and others regard them as fragments of olivine rock, that is, as xenoliths of dunite that were torn off from a dunite mass in depth and brought up by the rising basalt magma.

Exotic minerals. Some basalts contain unusual minerals, which do not appear in ordinary varieties and whose presence must therefore be ascribed to unusual conditions or to abnormal composition of the magma. The quartz basalt already mentioned is one of these uncommon varieties. Another is the iron-bearing basalt of Greenland, which contains small to large masses of native iron, much like the iron in meteorites. By the use of a solution of copper sulfate, specks of native iron have been identified as occurring in basalts from other places. The Greenland basalt contains graphite also.

Weathering and alteration. In many volcanic regions, basalts have been subjected to exhalations of steam or to hot water, and the minerals containing ferrous iron, such as magnetite and olivine, have become reddened through the change of their iron to ferric oxide. In some basalts the olivines alone have reddened; in others the whole rock has become deep red to reddish brown. Such rocks may be difficult to distinguish in the field from red felsites. However, the association with other rocks, the retained form of phenocrysts, and the good amygdaloidal structure, rarely seen in the most common felsites, may help one to recognize what the rock was originally.

The normal weathering of basalt in a temperate, humid climate gives rise to a ferruginous clay. In a hot monsoon climate marked by sharply defined wet and dry seasons, the ultimate product of weathering of basalt, as of other rocks (p. 179), is *laterite,* the residue from which all but the hydroxides of iron and aluminum have been leached.

When subjected to metamorphism, basalts react like the gabbros and dolerites previously described, and give rise to "greenstone" and to greenstone schists and amphibolite.

Occurrence of basalt. As sills and dikes, basalts of various types, both nonporphyritic and porphyritic, are common in all regions where igneous rocks occur. As lava flows they are vastly more abundant. There is scarcely a volcanic region in the world in which they do not occur in greater or lesser amount. In some regions, as in the lava fields of the Columbia Plateau in western America and of the Deccan traps in western India, they were poured out in stupendous masses, so that tracts of country 200,000 square miles in extent were flooded thousands of feet deep. A similar great field existed in northern Great Britain, and its remnants, portions still saved from the eroding edge of the Atlantic, form the greater part of the northern British Isles.

Leucite rocks. Basaltic rocks containing the feldspathoid minerals, nepheline or leucite, either accompanying the feldspar or taking the place of it, although not common, are fairly abundant in some regions. Ordinarily the feldspathoids are in the groundmass, to be detected only by means of the microscope, and such rocks would in the field be classed as basalts. In central Italy, however, leucite-bearing rocks are abundant and leucite crystals occur as phenocrysts as large as peas, or larger, so that they are easily recognized. They are leucite basalt porphyries, or leucite melaphyres. For the properties of leucite, its description under rock minerals should be consulted. According as other minerals are present, several different types of leucitic rocks are distinguished and named. Some are so light colored that they might be classed as varieties of felsophyres. Outside Italy leucitic rocks are rare, but they are known to occur in the Rhine district, in central Montana, western Wyoming, and especially in Java and central Africa.

GLASSY ROCKS

Many felsophyres and basalts are seen in thin section by the use of the microscope to contain glass, which acts as a bonding material to hold the mineral grains together. This glass cannot be detected megascopically, and the term glassy rocks, as here used, applies only to those that are wholly composed of glass or, if partly crystalline, contain enough glass to be visible to the unaided eye.

The conditions that cause a magma to solidify as a glass are evidently those unfavorable to crystallization, *i.e.*, extremely quick cooling and the rapid loss of its dissolved gas. This matter has already been discussed in connection with the texture of igneous rocks. These conditions prevail when magmas are poured out on the Earth's surface as lavas; moreover, just as we associate a coarse-grained, entirely crystalline, granular rock, such as granite, with an intrusive, deep-seated origin, so conversely we associate glassy rocks with an extrusive origin. Although some dikes and sills have glassy selvages along their borders because they were injected into cold rocks or consist wholly of glass because they were injected near the Earth's surface, as in dikes in volcanic regions, such glassy development is uncommon.

Any magma, no matter what its composition, will form a glassy rock if it is chilled fast enough. Glassy forms of felsites are common, but glassy forms of basalt are rare and of inconsiderable volume. The reason for this marked difference is that a felsitic magma has a high viscosity in the temperature range in which it begins to solidify, and as the magma cools down and approaches its solidifying range it becomes so enormously viscous that the free movement of ions necessary for crystallization is prevented. This extreme viscosity is due to the large amount of silica that such magmas contain, which has a strong effect in increasing the internal friction. The presence of water in the magma tends to counteract the effect of the silica and to make the magma more fluid and thus to help crystallization, but when the magma is poured out on the surface it loses its dissolved water rapidly, with resultant increase in viscosity. On the other hand, basaltic magmas, which are relatively low in silica and high in iron and magnesia, are much less viscous as they approach their solidifying range, thus permitting the newly born minerals to grow rapidly and the resulting rock to take on a stony texture and appearance. Accordingly, glasses high in silica and corresponding to granite in composition are those that are most common.

Classification of glassy rocks. The glassy rocks fall into two groups: (1) those containing phenocrysts embedded in glass, in short, porphyritic glasses; and (2) those without phenocrysts and consisting wholly of glass or of glass more or less filled with spherulites or lithophysae, as described later. The second group is subdivided according to luster and structure. Accordingly, we have the following classification:

Glass containing few or no phenocrysts	*Obsidian,* strong, bright, vitreous luster.
	Pitchstone, dull, pitchy, or resinous luster.
	Perlite, apparently made of small spheroids.
	Pumice, cellular structure, glass froth.

Glass more or less filled with phenocrysts	*Vitrophyre,* glass porphyry.

Obsidian. Obsidian is solid, natural glass, devoid of all crystal grains, or nearly so. It is brilliantly lustrous like artificial glass. It is generally jet black; but when the edges of thin chips are examined against the light they are seen to be transparent or translucent and to have a more or less smoky color. The coloring matter is more or less collected into fine parallel streaks, bands, or threads, having been drawn out during flowage. Less commonly obsidian is gray, Indian red, or rich brown, and these colors may be mixed with black in bands and strings, which, kneaded through it, produce a marbled effect. The black glass as shown by the microscope is colorless and filled with tiny, black, dustlike particles, which are probably specks of magnetite. These minute specks represent crystals that had just begun to grow; dispersed through the glass they act as a pigment, coloring it black. In some obsidians they have been oxidized to hematite dust, and the glass is then red or brown.

Obsidian has a remarkable conchoidal fracture (illustrated in Fig. 2, page 22) due to its homogeneity and lack of structure. This quality made obsidian highly valued by primitive peoples, for it enabled them by chipping to work the obsidian into desired forms—knives, spearheads, and other implements and weapons, including long slender flakes that for cutting purposes had knife-edges of razor-like keenness. The ancient Mexicans were especially skillful in working obsidian, and were able to spring off blades of bayonet-like cross section, ½ inch wide and 6 inches or more long.

Although obsidians corresponding to the various kinds of silicic igneous rocks are known, most of them have a composition like that of granite, as shown by the analysis by E. S. Shepherd, of a typical specimen from Obsidian Cliff, Yellowstone National Park.

SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O+	XyO[1]	Total
76.8	12.1	0.6	0.8	0.1	0.6	3.8	4.9	0.1	0.6	100.4

[1] XyO represents small amounts of other oxides.

It can be readily shown by calculation that this magma, had it become coarsely crystalline, would have produced a rock consisting of 35 per cent of quartz, 63 per cent of feldspar, and 2 per cent of other minerals, that is to say, a granite.

The specific gravity of obsidian ranges from 2.3 to 2.7, depending on the composition; that of the most common varieties is 2.3 to 2.4.

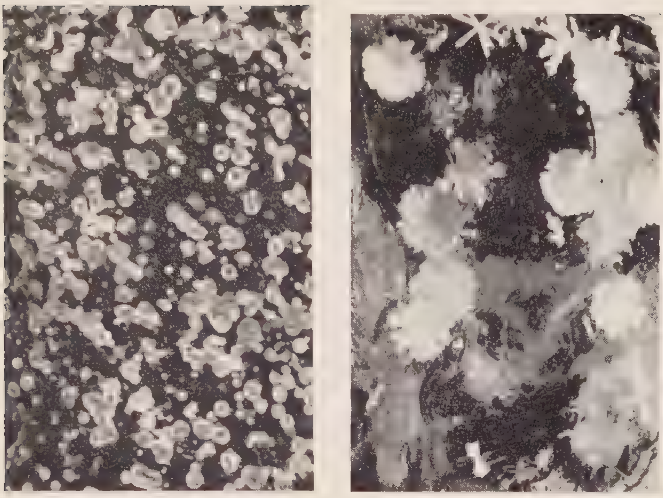
Its hardness exceeds that of ordinary window glass, which it scratches. Before the blowpipe a splinter of black obsidian fuses readily, accompanied by bubbling, to a vesicular gray or white enamel, which, after its water content has thus been driven off, is exceedingly infusible. This experiment is very instructive in showing the effect of water in lowering the fusing point of magmas and in increasing their liquidity. The water content of brilliantly lustrous obsidians is invariably under 1 per cent; it is believed to be magmatic water trapped in them as the result of rapid solidification. It is present in what the microscope reveals to be glass of the clearest kind, and it is not contained in cavities; it is a part of the obsidian, just as are the soda and potash.

Spherulites. Many obsidians contain round or even perfectly spherical bodies of white, gray, or red color, ranging from microscopic dimensions up to an egg or larger, but generally from the size of fine shot to that of peas. Such a spherical body if split open and closely examined with a lens can generally be seen to be composed of fibers that radiate from a center and to have also a concentric banding indicated by small differences in colors in the successive bands; the radiate body appears as if built of successive concentric shells. These bodies are called *spherulites* and consist chiefly of fibers of feldspar and interstitial crystalline silica. They indicate very rapid crystallization, during which the fibers grew outward from the centers where they began to crystallize and continued until stopped by the enormously increasing viscosity of the rapidly cooling magma. Spherulites should not be confused with phenocrysts, which are single individuals. An example is shown in Plate 22. Spherulites are sometimes formed by accident in artificial glass; in Plate 22 are shown such spherulites, which were formed in a flow of glass that broke out of a glass furnace.

The spherulites in many lavas are localized along flowage and fracture planes, so that on fracture surfaces at right angles to these planes they appear to be dotted along in lines. If the spherulites are numerous, they may coalesce; as a result some streaks in the obsidian consist of closely crowded spherulites, whereas other streaks or bands consist of the normal dark glass, as shown in Plate 20.

Lithophysae. Associated with spherulites in many glassy rocks are remarkable forms known as *lithophysae* (meaning "stone bubbles"). They consist of concentric shells of crystalline material that surround a central cavity and are more or less separated from each other. The shells consist of adherent crystals and are fragile. When exposed by the breaking of the rock, lithophysae resemble a full-blown rose with its concentric layers of petals. They range in size from very small to several inches in diameter. The walls of the cavities are coated with

PLATE 22.



A. Spherulites in Obsidian. B. Spherulites in Artificial Glass.

C. Lithophysæ in Rhyolite.

SPHERULITES AND LITHOPHYSÆ IN GLASSY ROCKS.

minute but beautiful crystals of quartz, tridymite, and feldspar, and less commonly, fayalite, topaz, garnet, and tourmaline. Some lithophysae are more or less flattened and are strung along the flowage planes of the rock. They occur not only in pure glassy lavas but also in lavas that are more or less crystalline. They are illustrated in Plate 22.

The origin of lithophysae is ascribed to rhythmic crystallization, each shell being the result of a single pulse of crystallization. The gases liberated during crystallization push apart the walls of the cavities by their expansive power. The formation of topaz, $Al_2SiO_4(F, OH)_2$, and other fluorine-bearing minerals points to the presence of fluorine in the liberated gases. Thus the lithophysae bear an analogy to miarolitic cavities in intrusive rocks (page 131), in that they were formed by the activity of gases liberated from the magma as a result of crystallization.

Pitchstone. Pitchstone is a variety of obsidian whose luster, instead of being bright and glassy, is resinous or pitchlike. It differs also in composition in that, whereas obsidian contains rarely as much as 1 per cent of water and generally much less, pitchstone contains much more, 5 or 6 per cent or even more. This high content of water probably gives pitchstone its distinctive luster. Pitchstones, like obsidians, are of many colors—black, gray, red, brown, and green—and many are varicolored; despite their dark colors they are translucent to transparent on thin edges.

Perlite. Perlite is a peculiar variety of glassy rock composed of small spheroids, ranging in size from small shot to peas. A well-developed botryoidal structure on fractured surfaces is therefore characteristic. Perlite is generally gray or blue gray, rarely red; it has a soft, pearly, or waxlike luster and resembles enamel. The spheroids either lie separated in a sort of matrix and are round, or they are closely aggregated and polygonal. They tend to have a concentric, shelly structure, which is a contraction effect in the cooling glass, the result of spherical, spiral cracking, as determined by study of thin sections.

Perlites prove on analysis to contain a rather constant percentage of combined water, between 3 and 4 per cent. There may be a connection between this amount of water and the peculiar method of cracking. To the casual observer they resemble the oolites and pisolites of the sedimentary rocks. Perlite occurs only in the highly silicic glasses; it does not occur in basaltic glasses.

Pumice. Pumice is extremely vesicular glass, produced by the expansion of superheated steam liberated at high temperature, as the result of decreased pressure on the magma as it came to the Earth's surface. Pumice is best described as glass froth. It is white, gray, yellowish, or brownish; rarely it is red. Some has a silky luster. Examined with the lens, it is seen to be a mass of silky fibers of glass, full of pores, and separated by larger holes like a sponge. If the fibers have been drawn out by flowage, they are parallel; otherwise they are intertwined.

The chemical composition of typical pumice is like that of highly silicic obsidian.

Pumice does not form independent rock masses; it occurs as the upper crust of flows of felsite lava, or in fragments among the material blown out by volcanoes. Because of its light, porous nature and its content of sealed glass cells, it floats almost indefinitely on water. Consequently the pumice ejected by volcanoes near or in the sea is borne by currents all over the world and drifts ashore everywhere. The value of pumice as an abrasive and polishing agent depends on the sharp cutting edges of the thin films and fibers of glass. Nearly the entire commercial output comes from the Lipari Islands, north of Sicily.

Scoriae. Although all magmas, whatever their chemical composition, may under the requisite conditions form pumiceous rocks, typical pumice, as just stated, is most characteristic of the felsitic magmas, and basaltic pumices are far less abundant. Nevertheless, basaltic magmas develop vesicular forms, by the expansion of their contained gases, as described under basalt. These vesicular portions, especially the tops of basalt flows, grade into slaglike modifications containing so many gas cavities of highly irregular shape and large size that there is at least as much empty space as solid matter; such material is said to be *scoriaceous*. Loose pieces of scoriaceous rock are *scoriae* (Plate 23).

A peculiar modification of what can be considered to be basaltic pumice occurs in the crater of Kilauea in Hawaii, where drops of lava flying up from the boiling lava lakes pull out after them thin, hairlike threads of glass. These threads, drifted by the wind, collect in towlike masses, called "Pele's hair," after the Fire goddess associated in native myths with the crater of Kilauea.

Vitrophyre. Obsidian and pitchstone may contain imbedded crystals, or phenocrysts. As in felsite porphyries, the phenocrysts range greatly in number: from few and widely scattered to many and closely crowded. Such porphyries, consisting of glass matrix and phenocrysts, are called *vitrophyres*. Perlite porphyries are known but are rare.

The glass matrix of the vitrophyres has the properties of the obsidian or pitchstone previously described; it may contain spherulites in addition to the phenocrysts. Feldspar is the mineral most common as phenocrysts; it is generally limpid and glassy, but its cleavage distinguishes it from quartz, which also may occur, alone or with the feldspar. Mafic phenocrysts, if present, are generally biotite and less commonly hornblende; pyroxene is rare.

Vitrophyres are named according to the phenocrysts they contain, without regard to the composition of the groundmass; thus we have *quartz vitrophyre, feldspar vitrophyre, quartz-biotite vitrophyre,* etc. The phenocrysts are generally rather small.

Vitrophyres are similar in chemical composition on the one hand to felsites and on the other to pure glasses. They represent an intermediate stage of crystalline development, as may be seen from Table 7, which shows the relations of these varieties of lavas to one another.

TABLE 7

INTERRELATION OF SILICIC LAVAS AND MAGMAS

Conditions under Which Magma Cooled and Solidified	No crystals formed in the depths before extrusion; no phenocrysts	Crystals formed in depth before extrusion and brought up by magma; has phenocrysts
No minerals crystallized from magna on Earth's surface, because of rapid cooling. Glassy texture.	Pitchstone and Obsidian	Vitrophyre
Minerals have crystallized from magma on Earth's surface; slower cooling. Stony texture.	Felsite	Felsophyre

Tachylyte. Basaltic magmas readily become crystalline and rarely form glass and only in relatively small volume. Basaltic glass occurs as thin selvages in dikes, as thin crusts on lava flows, or among the fragmental products of basaltic volcanoes. It is known as *tachylyte* (Greek, *tachy,* quick, and *lysos*, soluble, in reference to its ready solubility in acids).

Occurrence of glassy rocks. Glasses occur in regions that are or once were scenes of volcanic activity. Although obsidian and pitchstone occur as independent flows and masses near volcanic vents, they form as a rule only the tops of lava sheets, which are crystalline lower down. Glassy rocks, especially the pumiceous forms, occur also in the fragmental material ejected by volcanoes. To attempt to name all the occurrences is impracticable, but a few of the more notable may be mentioned. Obsidian occurs in large masses in Yellowstone National Park, where it is known for its beautiful spherulites and lithophysae, at Mono Lake in California, and at many other places in the United States and elsewhere in the world. Pitchstone occurs near Georgetown and at Silver Cliff, Colorado.

Alteration of glassy rocks. Many ancient lavas were once glassy, though they are not so now. When a natural glass is exposed to the various agents that tend to alter rocks, such as heat, solutions, and shearing stress, it is converted into an intimate mixture of extremely minute grains of quartz and feldspar and thereby loses its glassiness. It takes on a stony texture and becomes an aphanitic felsite. This change is *devitrification.* That many felsites are devitrified glasses cannot be proved, even by means of the microscope, but it can be suspected from the presence in them of chains of spherulites, flow structures, and lithophysae, which can be seen megascopically and strongly hint that the rocks in which they occur were formerly glassy. Ancient devitrified lavas of this kind were first recognized in America at South Mountain, Pennsylvania. Devitrified felsites closely resemble the sedimentary rocks called cherts, but on critical examination in the field, particularly of their weather-etched surfaces, they can generally be recognized to be of volcanic origin.

FRAGMENTAL VOLCANIC ROCKS

Origin. Fragmental igneous rocks represent the material that was blown out by volcanoes during explosive activity. The explosive action is due to gases, chiefly superheated steam, contained under pressure in the magma, and as the magma rises to the surface and the pressure on it is relieved, its dissolved gases escape with violence. While the major part of the gas content of the magma is escaping in great volumes, which rush upward and carry the solid or liquid materials to great heights, a minor part is expanding within the liquid, inflating it into vesicular forms. Consequently the solidified particles as they fall are commonly spongiform. Mixed with them however, are compact pieces of lava and other rocks—fragments of crusts of solid lava formed by cooling after a previous eruption, and pieces torn from the rock walls of the conduit. After the gases have mostly escaped, explosive activity ceases, and the violent ejection of material is likely to be succeeded by quiet extrusion of liquid rock. Hence beds of fragmental material commonly alternate with layers of compact lava—felsite or basalt.

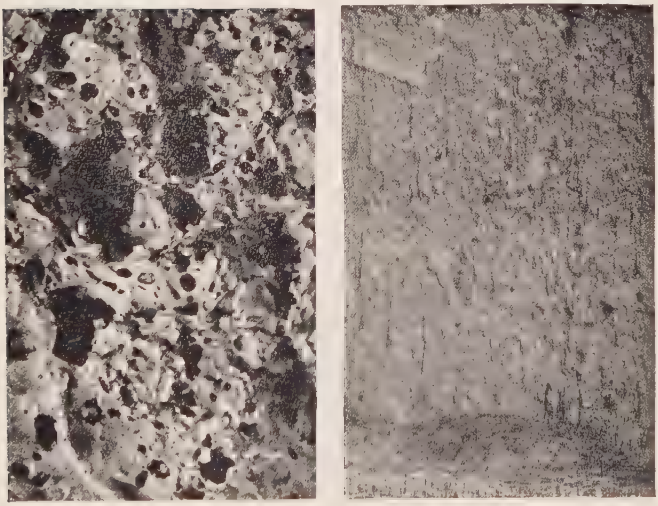
The chemical composition of the nonvolatile portion of the magma also plays a part in determining explosive activity. The magmas corresponding to felsite and therefore high in silica are, as already mentioned, extraordinarily viscous at temperatures at which those low in silica, such as basalt magmas, which are rich in iron, magnesia, and lime, are still relatively very liquid. From the stiffly viscous silicic magmas the gases escape with difficulty and with explosive violence; from the basaltic magmas they escape readily and nonexplosively. Although exceptions to this generalization are many, it is substantially true. Therefore, most vents from which felsophyre lavas were erupted are surrounded by high and steep cones, composed chiefly of fragmental material, whereas basaltic vents are surrounded largely by outflows of lava, and the resulting volcanic edifices are therefore broad and relatively low in comparison with their great breadth. Many volcanoes, Vesuvius for example, are of intermediate character: explosion and ejection of material alternate with outpouring of lava, and the resulting edifice is of composite construction.

Classification. The particles of magma blown into the atmosphere and solidified and the ejected pieces of old rocks range in size from the finest dust, which may float for years, to huge masses weighing hundreds of pounds or even many tons. For convenience, the following sizes are distinguished: pieces larger than 32 millimeters in diameter are called *bombs* if ejected while still viscous and *blocks* if ejected as solid pieces; those between 32 millimeters and 4 millimeters are termed *lapilli;* and those smaller than 4 millimeters, *ash;* the very finest material, *volcanic dust,* is included with ash. Most blocks and lapilli are sharply angular, but some are smoothly rounded, the rounding being caused by the grinding and attrition of the pieces upon one another in the upward rush from the volcanic throat. Some of the larger bombs are of subangular shape, are porous, and have their surfaces disrupted by cracks, as shown in Plate 23; they are well named *bread-crust bombs.*

In the course of time an enormous amount of material blown out from volcanoes, *pyroclastic material,* may be deposited over a vast territory, covering up existing rock formations. The heavier and coarser particles fall first and then the finer, the result being a gradation from bottom to top called *graded bedding*; as successive outbursts follow one another, a rough bedding is produced. By its own weight as it accumulates, aided by the action of percolating water which may deposit some of its dissolved substances in the interstices between the particles, the pyroclastic material gradually becomes compacted into a more or less firm mass having a certain individuality as a kind of rock. If the resulting rock is composed entirely of the finer particles, dust and ash, it is called *tuff*; if made up dominantly

PLATE 23.

A. VOLCANIC BOMB.



B. SCORIA.

C. PUMICE.

of lapilli it is called *lapilli tuff*; and if made up of blocks and bombs, it is called *agglomerate.* Customarily, however, pyroclastic rocks composed largely of the coarser fragments, are called *volcanic breccia*, from the angular character of the embedded fragments.

Tuff. Tuffs by definition are fine-grained rocks. Many are of a chalky consistency, compact, and breaking into small chips. They are light in color—white, pink, pale brown, gray, or yellow—rarely passing into darker shades. Indurated tuffs can easily be mistaken for felsite lavas; some, in fact, cannot be distinguished from felsite megascopically, but on close examination with a good lens angular particles of quartz, feldspar, and other minerals can generally be seen in them, and possibly small fragments of other rocks. If not too firmly indurated, tuffs have a rough feel, and when strongly rubbed between the fingers they yield a gritty dust, unlike the smoothness of pure clay or chalk, the gritty feel being caused by the hard, angular character of the dust particles.

Some tuffs contain fossil remains of vegetation, if the ash fell on land surfaces covered with vegetation, and they may contain carbonaceous remains of stems, twigs, or leaf imprints. If the ash fell into water, the resulting tuff may be rich in fossils of various kinds, such as marine organisms, possibly of great perfection of form, and for the same reason it may be well stratified.

Volcanic breccia. Breccia has a cement or bonding of tuff, more or less completely filled with angular lapilli, together with still larger material—blocks and bombs. Interspersed with this "new" volcanic material are likely to be fragments of older rocks, such as limestones, shales, sandstones, and massive crystalline rocks, granite, gneiss, schist, and others—pieces of the foundation through which the conduit has been drilled. Tuffs, even after they have become greatly indurated by contact metamorphism or other agents, still reveal by differences of color and texture on a freshly broken face the angular shapes of the fragments and their composite character. If not too indurated, they are likely to be eroded very unevenly. The finer cement, being less resistant, washes away first, leaving the contained fragments projecting, and in this way, along the edges of cliffs, strange and weirdly shaped figures of erosion are produced. The color of breccias is varied: browns, reds, and chocolate are common, along with lighter tones, depending partly on the state of oxidation of the iron-bearing compounds they contain and partly on the nature of the rock type, whether felsophyric, which has light colors, or andesitic and basaltic, which are dark.

PLATE 24.

BEDDED VOLCANIC BRECCIAS, TABLE MOUNTAIN, ABSAROKA RANGE, WYOMING.

(J. P. Iddings, U. S. Geological Survey.)

Occurrence of tuffs and breccias. Pyroclastic rocks are widely distributed, occurring in all regions where volcanic activity has taken place. Their presence indeed is conclusive evidence that such activity took place in the past. Wherever volcanism is still active or has recently ceased, they are represented by uncompacted material.

In the Rocky Mountains, in the Coast and Cascade Ranges, and in fact over most of western North America, they are common rocks; and in many places in Colorado, Wyoming, and Montana they occur in immense deposits, forming much of the bulk of the mountain masses. They are especially well displayed in Wyoming, in the region east of Yellowstone Park, where the serried peaks of the Absaroka Range are carved mostly out of tuffs and breccias that aggregate 11,000 feet in thickness, eloquent testimony to the enormous explosive volcanic activity that this region formerly displayed. Sections cut by erosion into this great pile of pyroclastic rocks reveal their rough bedding, as shown in Plate 24. In this region they contain many intercalated flows of lava.

Tuffs and breccias are interbedded at many places with ancient sedimentary rocks, and they may have been so greatly altered by metamorphic processes that they can be recognized only by careful petrographic research. Altered tuffs form much of the oldest Precambrian formation of Sweden and thus comprise some of the most ancient rocks of our planet. Tuffs and breccias are widely distributed throughout the world.

CHAPTER IX

ORIGIN AND CLASSIFICATION OF SEDIMENTARY ROCKS

The sedimentary rocks consist of material that was derived from the destruction of previously existing rocks. This material was transported from its source and deposited by the action of water, wind, or glacial ice and was later converted into rock. The activity of water is by far the most important, because of the great volume of rocks that have thus been formed. Accordingly, when sedimentary rocks are mentioned, water-laid rocks are meant unless it is otherwise stated. In contradistinction to them, the material that was moved and deposited by wind forms the class of *eolian rocks*, which are far less abundant; ice-laid rocks are still less abundant.

From what has been said, manifestly, the sedimentary rocks are secondary, in that the material of which they are made was derived from rocks already in existence. This material has resulted from the reworking of former rocks of all classes—igneous, metamorphic, and sedimentary—and, in the earliest sediments, from the Earth's original crust, if such ever existed.

The rocks formed in bodies of water are divided into two main groups, according to the way in which the material was deposited: they consist (1) of detritus that was transported mechanically and on coming to rest sank as a sediment; and (2) substances that were carried in solution, were separated from solution by chemical and physiologic processes, and were then deposited. The first group we shall call *detrital*, the second *nondetrital* sediments. Yet even between these two groups, as we shall see later, it is difficult to draw a definite line.

Decay of rocks; formation of soil. When rocks are exposed to the action of the atmosphere, they gradually decay and eventually turn into soil. This result is brought about by the cooperative action of many factors. All rock masses are penetrated by cracks and by joints. These passageways are both great and small. The individual mineral grains in addition contain cleavage and other cracks. Water is therefore able to penetrate and permeate the rock masses. In cold regions where thawing and freezing alternate, the expansion of the water in turning to ice keeps splitting the rocks near the Earth's surface until they are reduced to a mass of debris. The expansion and contraction of rocks in hot countries, under great daily and yearly changes of temperature, accomplishes the same thing more slowly. The expansive power of growing roots of trees and plants tends to the same end. By such processes the rock masses tend to be broken up, *mechanically*, into smaller and smaller fragments.

In the meantime the substances dissolved in the water, such as oxygen and carbon dioxide from air, and acids, including more carbon dioxide from the decaying vegetation, are acting *chemically* on the rock minerals. They convert the silicates, oxides, and sulfides into new compounds, such as carbonates, hydrated silicates, hydroxides, and sulfates, some of which are soluble. The soluble material is dissolved and is carried away by running water into streams and eventually into the ocean, where it accumulates in increasing concentration as a result of evaporation of the water. Some minerals, such as quartz, are not attacked under ordinary circumstances to any appreciable extent, or but very slowly, and they remain as the chief part of the rock debris. For this reason the resistant minerals, especially quartz, predominate in the sedimentary rocks that are formed by mechanical processes. This mantle of broken and altered rock is called by various names, and the finer upper portion in which vegetation grows is the *soil*. Under this name, for convenience, we can consider all of the mantle of weathered and partly weathered rock. The gradual transition from unaltered rock below to soil above is illustrated in Plate 17. In soil science such a complete vertical section extending down to the parent material is called a *soil profile*.

Movement of soil. The surface of the land in general is covered by a mantle of soil resting on the rocky crust of the Earth. Here and there in ledges, precipices, and the craggy tops of hills and mountains, the rocks project through this covering. By the action of running water, aided by gravity, this altered rock and overlying soil, although it is apparently at rest, is creeping continuously downward toward the streams and thus to the sea, its ultimate goal (see Plate 25). On steep slopes it moves more rapidly, in valleys more slowly; in level plains it is temporarily impounded like water in a lake. Its rate of motion varies from time to time and from place to place. Its movement in mass is, of course, very slow, but after it enters the streams, its rate approaches that of the water that carries it; when it is resting on the stream bottom, the rate varies according to circumstances. Thus the rock waste is ever being carried away and ever renewed by the destruction of the underlying rocks. The greater part of the rock

PLATE 25.

TALUS CONES, SHOWING MOVEMENT OF ROCK WASTE; HEAD OF SHOSHONE RIVER, WYOMING.

(U. S. Geological Survey.)

waste is carried into the ocean, but much is deposited in lakes, inland basins, and on the flood plains and deltas of rivers. The material thus laid down in these diverse environments comprises the sedimentary deposits of mechanical deposition. The rock waste in transit, as already mentioned, is called *detritus*; hence the sediments formed from it are *detrital sediments* and the rocks eventually resulting from these sediments are *detrital rocks*.

Classification of sedimentary particles by size. Detritus resulting from the erosional destruction of land masses consists of materials of greatly differing sizes. The range is particularly evident in glacial detritus, because glaciers move rock fragments differing enormously in size and dump them in chaotic admixture. When detritus is moved by running water, it becomes sorted, or graded, according to the strength of the current, into masses consisting of approximately equal-sized particles. If the deposit consists of particles larger than about 2 millimeters, it is called *gravel* and the individual pieces are called *stones*, or better, *roundstones*. Large roundstones exceeding 256 millimeters (about 10 inches) in diameter are spoken of as *boulders*; those between 256 and 64 millimeters in diameter are *cobbles;* and those between 64 and 2 millimeters in diameter are *pebbles*. This rather arbitrary classification, which however has the advantage of precision, has come into general use in recent years. Particles smaller than 2 millimeters in diameter, which form an incoherent mass when wet with water, are called *sand grains*. Still finer particles make up the deposits known as *silt*; on drying after being wet silt particles generally cohere into solid material. The very finest detritus—muds and clays—is made up of particles of colloidal dimensions, less than 0.002 millimeter (2 microns) in diameter. Table 1 gives the classification according to size.

TABLE 1

GRADE SIZES OF SEDIMENTARY PARTICLES

Name of Material	*Diameter in Millimeters*
Boulders	256 or more
Cobbles	256–64
Pebbles	64–2
Sand grains	2–0.02
Silt grains	0.02–0.002
Clay particles	0.002 or less

Gravel. Gravel is a sediment consisting of rounded fragments predominantly larger than 2 millimeters in diameter and generally containing admixed sand. The rounded fragments—pebbles, cobbles, or boulders, depending on the coarseness of the gravel—are mainly pieces of individual rocks and, like them, are generally made up of grains of various minerals. Some pebbles and cobbles are composed of only one mineral, quartz being by far the most common. Such quartz pebbles are fragments that were derived from quartzite strata, from quartz veins, or from large quartz crystals in granite pegmatite. Coarse granites and pegmatites may furnish other pebbles consisting of a single mineral, especially feldspar.

The shape and appearance of sedimentary particles depend on many factors. Particles that were rolled long distances in the beds of streams or the shores of lakes and of the sea are well rounded, ranging in shape from ellipsoidal to spherical. They are likely to have a smooth surface of a characteristic faintly dimpled, slightly dented appearance, caused by repeated collisions during transport. This feature is best seen on a pebble of a hard homogeneous substance such as quartz. If the pebbles are composite in mineral composition, they are generally pitted as the result of the removal of the softer or more easily altered minerals.

The degree of rounding of sedimentary particles depends on the distance and length of time they have been transported and on the tenacity and hardness of the material. Some sedimentary rocks, as will be shown, are composed of pebble-sized fragments that did not travel far and therefore still retain their original rough, angular shapes.

Rock fragments that were transported by glaciers occur in some sedimentary rocks. Some of these fragments have characteristic subangular forms, with facets ground on them, which are polished and scratched by parallel and intersecting grooves and striae. Pebbles partly buried in the sand of the seashore and of deserts may become subangular and faceted, the facets having been ground by the sand drifting past them, but they lack the scratches. Characteristically they approximate in shape a butternut, or Brazil nut. Such wind-fashioned stones are termed *ventifacts*.

Pebbles buried in the soil may show fernlike or mosslike markings (termed dendrites) upon them, or they may have a shiny skin of dark color. This comes from manganese or iron oxides that are deposited by water.

Sand. Sand, strictly defined, means sedimentary particles of a certain size (between 0.02 and 2 millimeters, as just mentioned). It implies nothing in regard to composition. Thus we have quartz sand, coral sand, and volcanic sand. However, most sands consist of particles of quartz, and some are wholly made up of quartz. For this reason, "sand," when used without qualification, is taken to mean quartz sand.

Sand is variable in composition. Quartz generally dominates, but many other minerals may be present, depending on the rocks in the region from which the sand was derived. Feldspar is by far the most abundant after quartz. Garnet and magnetite are common. Hornblende, pyroxene, and tourmaline are also likely to occur. Some grains consist of pieces of very fine-grained rocks of composite character. Twenty-three different minerals occur in the dune sand of Holland, but the total amount of these minerals other than the quartz is only a fraction of one per cent.

Sand grains are more or less rounded, depending on a number of factors: their size, their specific gravity, the distance they have been transported, and the transport medium. In some sea sands they are almost all spherical. Below a certain size—about 0.1 millimeter—sand grains do not become more rounded by attrition in water, because the capillary film of water surrounding the grains acts as a buffer and prevents them from coming in contact when they collide; however, for the larger grains the buffering action is ineffective. In wind-blown sands, especially those of the desert, the quartz grains tend to become rounded to minute perfect spheres having dull, mat surfaces like ground glass. In immature desert sands, *i.e.*, those not far from their source rocks, the grains may of course be quite angular.

Silt, mud, and clay. Silt and mud are sediments composed of the very finest-grained products of erosion. *Silt* is so fine grained that, unlike sand, it coheres when wet. *Mud* and its principal variety, *clay*, consist of particles that are smaller than those of silt size, being less than 0.002 millimeter in diameter. As sedimentary deposits muds are characteristically found off shore, or in sheltered bays and sounds, where water moves too slowly to transport the heavier sand and gravel, and in the material forming the lower flood plains and deltas of rivers. Because of their minuteness the particles are not likely to have become rounded, and in fact under the microscope they are seen to be angular. Like the sands, they are composed of a great variety of minerals—kaolinite, mica (muscovite and bleached biotite), quartz, feldspar, and others. Just as quartz is the characteristic mineral of sands, so kaolinite and allied montmorillonite, illite, and halloysite, the so-called "clay minerals," are characteristic of clays. As shown previously, the feldspars of the rocks decay to produce kaolinite and other flaky clay minerals, whereas the quartz remains unaltered. The flaky minerals tend to float because their surface is large as compared to their weight, but the quartz grains tend to sink. Because of this difference the quartz tends to become separated from the flaky minerals by moving water: as the current slackens, the quartz is deposited first, forming sand, while the finely flaky clay minerals are carried beyond and settle in still water. Minute flakes of white mica generally accompany them.

Clays, which, as just mentioned, consist of the very finest detrital particles, of "colloidal dimensions," tend to remain in suspension almost indefinitely in fresh water. However, if salt is added to such a suspension, or if, as in nature, the suspension becomes mixed with sea water, the clay curdles into lumps, "flocculates," and is quickly deposited, leaving the liquid clear. This behavior has an important bearing on the character of the resulting sediments, which depends on whether they were deposited in fresh water or in salt water: fresh-water clays are laminated and remarkably well stratified and show graded bedding, as exemplified by the annually layered glacial clays, whereas the corresponding marine clays are devoid of these features.

Muds are characterized according to their predominant constituent. Those containing a considerable amount of finely flaky minerals are called clays, or argillaceous sediments. Far less common are coral muds, consisting of finely triturated coral debris; aragonite muds, such as form on the Great Bahama Bank; volcanic muds; and others.

Origin of the material of the nondetrital sedimentary rocks. Erosion of the lands supplies not only material that is mechanically transported by the streams, but also material that is taken into solution and thus carried to the sea. According to Clarke, the dissolved material removed from the lands is 2,740,000,000 metric tons per year. The amount thus carried to the ocean implies that the land surface of the globe is being lowered by solvent erosion at the rate of one foot in 30,000 years. The dissolved matter differs greatly in chemical composition in different rivers, being determined by the composition of the rocks forming the drainage basins.

As the result of the continuous evaporation of water from the ocean, the soluble salts carried to it by the rivers of the world have steadily accumulated in it, and it has thus attained the high salinity it now has. Much of the dissolved matter brought to the ocean has been precipitated from time to time by the physiologic activities of living organisms and by physical-chemical processes. Thus were formed the nondetrital sedimentary rocks. These rocks include the immense volume of marine carbonates (limestones and dolomites), and the less abundant sulfates and chlorides, such as anhydrite, gypsum, and rock salt.

The carbonates of calcium and magnesium were ultimately derived from silicates of these metals. Water containing carbon dioxide converted the silicates into carbonates, as illustrated by the decomposition of feldspar, and then dissolved the carbonates and carried them into the sea. The sulfates were formed by oxidation of sulfides in the original rocks. The chlorides in the ocean, comprising 77 per cent of its total dissolved matter, were derived in part from minerals in igneous rocks, but almost wholly, it appears, from the primeval atmosphere at the time of the condensation of the globe and from volcanic emanations during the subsequent span of geologic time.

Structure of sedimentary rocks. The sedimentary rocks, as geologic masses, form widely extended, relatively thin, tabular bodies, making a coating upon the Earth's outer shell. Their most characteristic feature is that they are *stratified.* This means that they are built up of layers, one above the other, differing in material, texture, color, thickness, or all of these properties. An individual layer is called a *bed* in plain Anglo-Saxon, or a *stratum* (plural, *strata*) in technical language. If the beds have not been disturbed by geologic forces since they were laid down, they are horizontally disposed, one above the other. Horizontal beds are shown in Plate 26.

Volcanic ash deposits are rudely stratified, rarely well stratified. The heavier particles blown out from a volcanic vent fall first and are succeeded by smaller, lighter ones. Repetitions of this process build up a sequence of individual beds superposed one on the other and thus produce a rude stratification (see Plate 24), and the lighter dust may be deposited by air currents much as if it had been laid down in water.

To develop stratification during sedimentation, variation in the conditions of deposition or in the size of the particles is necessary. During a period in which uniform conditions prevail, stratification will not be developed. Thus the eolian deposit, loess, which consists of the finest dust laid down by the wind, is so uniform in the size of its constituent particles in any one area that it is without stratification. Deposits of calcium carbonate, such as limestone and chalk, may form in the open ocean under such uniform conditions and with such uniform grain size that beds of these rocks more than a hundred feet thick are quite structureless and devoid of stratification.

Parallelism of layers in a rock mass is not necessarily a mark of stratification and therefore does not prove that a layered rock is of sedimentary origin. The mineral composition, texture, and relation to the accompanying rock masses must also be taken into consideration. For example, flowing lava may draw out portions of itself into a series of thin superimposed sheets that differ from one another, as illustrated

PLATE 26.

ALTERNATING BEDS OF LIMESTONE AND SHALE, COLORADO.

(U. S. Geological Survey.)

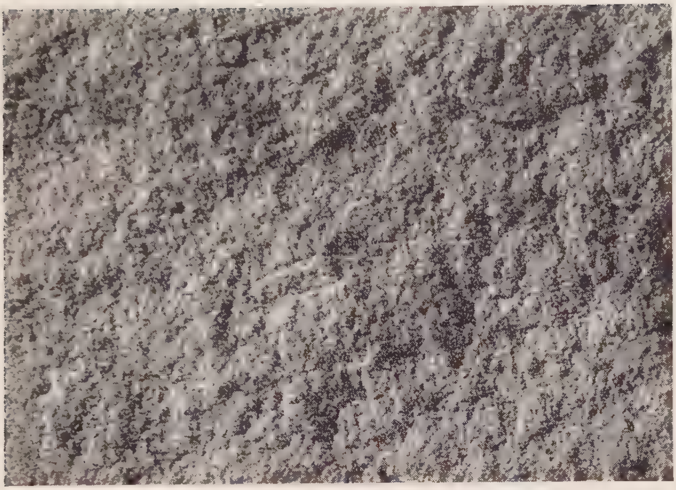
in Plate 20. Such flow-layered lavas are easily recognizable if fresh and unaltered, but if intercalated in sedimentary deposits and changed in appearance by geologic agents they may be confused with the accompanying stratified rocks. Again, perfect parallelism of layers and structure may be induced in rocks of all kinds by the shearing and metamorphism that accompany deep-seated crustal folding. The parallel structures induced in rocks by metamorphic processes may simulate stratification closely, but consideration of the points mentioned generally suffices to show the difference between them. Serious errors in understanding the real origin of the rocks may arise through failure to appreciate these facts.

The individual layers that are uniform in texture, color, and composition in a sequence of stratified rocks may range from the thinness of paper to a hundred feet or more in thickness. A layer that has a general uniformity of character and composition that serves to distinguish it clearly from other beds above and below it may be seen to be made up of smaller subdivisions whose differences from one another are not conspicuous. These smaller subdivisions are called *laminae.* A laminated sandstone is shown in Plate 27. As explained above, laminae are wanting in beds formed under uniform conditions. The individuality of a bed is generally emphasized by its particular hardness and appearance, its method of cracking or jointing, and the way in which it is affected by erosion, as compared with the same features of the beds above and below it.

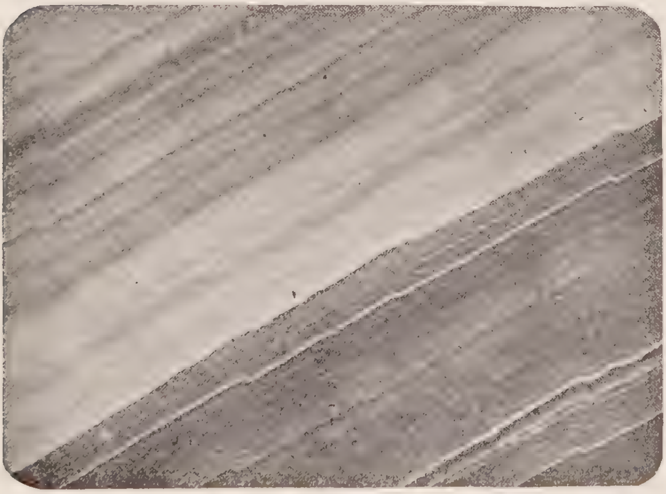
Texture of sedimentary rocks. The texture depends on the grain size of the particles or on their shape, and on the amount and character of the bonding material or cement, which determines the coherence of the rock. The grain size ranges within wide limits, but as explained previously under grading of material, grain size in itself largely determines the kind of rock. Thus conglomerates are by definition coarse-grained detrital rocks; sandstones are medium grained; and shales are fine grained. In each of these classes there is a range in grain-size: thus we have fine-, medium-, and coarse-grained sandstones; medium grain size in sandstone is about that of cube sugar.

The shape of the component grains depends on the past history of the grains, as explained under gravel. The grains may be more or less rounded, but most are broken, angular particles. The angular shape of the grains in many medium-grained sandstones and arkoses can be distinctly seen by close observation with a good lens. Rocks made up of such broken angular particles are said to have a clastic texture. In the coarser rocks the clastic character is particularly striking.

PLATE 27.



A. SANDSTONE, OF FINE GRAIN.



B. LAMINATED SANDSTONE, WITH SLIGHT FAULTS.

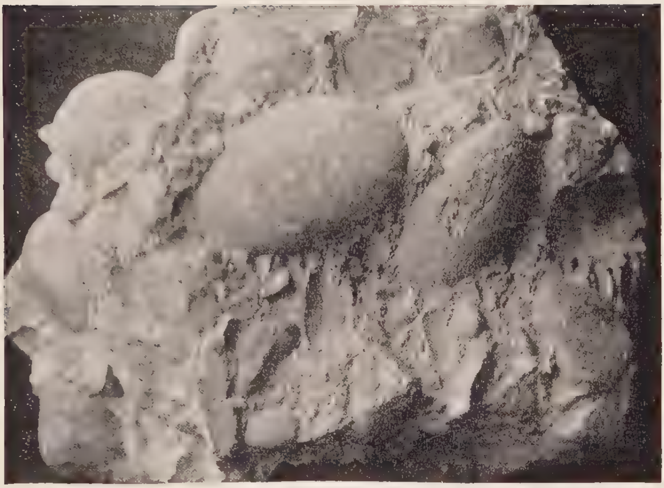
Rocks made up of large angular pieces are sedimentary breccias and and are called *fanglomerates*. These are illustrated in Plate 28. Such breccias, however, should not be confused with volcanic breccias, described on page 224.

The *cement* is the material that binds the clastic particles of sedimentary rocks together. It has bonded the loose material into a firm rock. Various substances act as cements. The cements in some rocks were brought in solution from outside sources and were deposited in the pores; in other rocks part of the sediment itself was dissolved and was redeposited; and in others the cement consists of fine material mechanically inclosed with the sediment. Silica and calcium carbonate are the common cements formed in the first and second ways; in the third, clay or claylike substances are the bonding material. Iron oxide is a common cement, in the form of hematite, goethite, and limonite. Many muds and clays have become consolidated into firm rocks apparently as a compaction effect brought about by the weight of the beds overlying them, without the presence of a perceptible cement.

The firmness of the rock resulting from the consolidation of a sediment depends, then, in part on the amount and nature of the cement, and in part to the pressure to which the rock has been subjected. As a result sedimentary rocks differ widely in coherence or firmness: some are very hard, firm, and compact, breaking like igneous rocks under the hammer and susceptible of taking a polish, such as some limestones and sandstones; others are so loose and incoherent that they can readily be rubbed to powder between the fingers, as can be done with chalks and some sandstones.

Color of sedimentary rocks. In describing a sedimentary rock, or any other rock, we mean when we speak of its color the color of the fresh, unweathered rock. This color may differ greatly from that of the weathered rock. It depends partly on the color of the constituent mineral particles and partly on the pigments disseminated through the rock. The most common constituents of the sedimentary rocks are quartz, kaolinite, feldspar, calcite, and dolomite; these are normally white or colorless, though some may have an exotic coloration. Rocks composed entirely of the normal minerals, without included pigment, are white, as illustrated by certain sandstones, clays, and chalk. More or less pigment is generally present, however, the common pigments being iron compounds and carbonaceous matter. The iron is in the form of ferric oxide, as hematite and probably hydrohematite (turgite), which give red to red-brown colors, and as limonite and goethite, which produce yellow to yellowish-brown

PLATE 28.



A. CONGLOMERATE, OF SEDIMENTARY ORIGIN.

B. BRECCIA, OF SEDIMENTARY ORIGIN; TRIASSIC, CONNECTICUT.

tones. Carbonaceous matter or finely divided carbon is black, and consequently rocks containing an abundance of such matter are black; as the amount lessens, dark grays are formed, and so on into pale grays. If both carbonaceous matter and iron oxides occur in a rock, the carbonaceous matter exerts a control on the coloring capacity of the iron in this way: in the presence of organic matter, especially when it is decaying, iron is reduced from the ferric to the ferrous state, changing from ferric oxide to ferrous carbonate; and as ferrous compounds are colorless or light colored, the rock has the tones of color produced by the carbonaceous pigment.

If such rocks are exposed to weathering, the carbonaceous material is destroyed, the iron becomes reoxidized, and red and yellow colors show themselves. This change is illustrated in outcrops and on joint faces of many black slates, which weather to red or yellow. On the other hand, if rocks are devoid of iron, they become white or light colored on weathered surfaces when the organic pigment is bleached out by the action of sunlight. Moreover, if solutions containing organic matter percolate through the rocks, the iron in ferric compounds is reduced to the ferrous state, and when thus reduced, or if in the ferrous state originally, it is dissolved and is carried away as the soluble ferrous bicarbonate, the rocks thus becoming light or colorless.

The prevalent colors of sedimentary rocks then are white through light gray to dark gray and black; white to pink, red, dark red, and red brown; and pale yellow to buff and yellow brown. The reds and yellows may be commingled in the same rock mass or layer, according to the varying distribution of the iron hydroxides. The colors of conglomerates and coarse arkoses are modified of course by the colors of the fragments of rocks and minerals in them.

Chemically formed rocks. Rocks formed by chemical means consist of material that has been precipitated from solution. There are three ways by which dissolved material is precipitated: (1) by evaporation; (2) by precipitants formed by the biochemical activity of living organisms; and (3) by the activity of living organisms in building their shells.

The deposits produced by evaporation are formed when bodies of sea water become partly or completely shut off from the ocean by geologic processes and become so highly concentrated by evaporation that they are unable to hold the salts in solution. The salts are then precipitated in the order of their insolubility. Precipitation thus effected is, strictly speaking, a physical process, not a chemical process. Gypsum and anhydrite (sulfates of calcium) and common salt (sodium chloride) are the most voluminous deposits laid down in this way. The rocks thus formed are appropriately called *evaporites*. Similar deposits form in arid regions in lakes and inland seas that have no outlet. The dissolved material brought by the inflowing streams is steadily concentrated by evaporation, and eventually it precipitates. Carbonates, sulfates, and chlorides are the main salts deposited.

Deposits are formed by biochemical reactions of various kinds. Ammonifying bacteria produce ammonia, which reacts with sea water to precipitate calcium carbonate. Certain algae are able to abstract from water the carbon dioxide they need, and the removal of the carbon dioxide causes the dissolved calcium carbonate to be precipitated.

Animals living in water, notably those living in the sea, secrete as a normal physiologic activity inorganic material to form their hard parts, either skeletons to stiffen them or shells as defensive armor. When the animals die, these hard parts accumulate as deposits. The chief substances secreted are calcium carbonate ($CaCO_3$) and opaline silica ($SiO_2 \cdot nH_2O$), the carbonate being much the more abundant. Examples are the reefs formed by corals and the shellbanks made of shells secreted by mollusks. Certain animal organisms (radiolaria and sponges) and microscopic unicellular plants (diatoms) secrete silica, giving rise to siliceous deposits. Diatoms are remarkable in contributing to the accumulating deposits not only as the result of perishing because of some natural disaster, but also during fission, when the adult diatom divides into two daughters and sheds its shell.

When water in moving through the rocks dissolves substances and later attains the outer air, as in springs, part of the dissolved load is deposited. The deposits thus formed, geologically considered, are of minor volume. They are illustrated by the deposits of travertine around springs and by the dripstone in caves, and by the siliceous sinter around the orifices of geysers and hot springs in volcanic regions. Other deposits are those formed where water has leached iron from the rocks and soils and, carrying the dissolved iron into swamps and shallow waters, has deposited it there, as ferrous carbonate (siderite) if abundant organic matter was present, or in reoxidized form of ferric hydroxide (limonite) if organic matter was wanting.

Circulation of material of the Earth's crust. Geology is not yet able to make definite statements on the origin of the material of the earliest-formed sediments on the Earth. Wherever erosion has exposed the foundation on which the very oldest sediments were deposited, the foundation is seen to consist of igneous rock or of igneous rock that has been metamorphosed, and the sediments are such as could have been derived from the weathering and erosion of the foundation rocks. Whatever the nature of the first sediments was, manifestly all later sediments have been derived only from the following sources: (1) from the first-formed sediments after they had been elevated to form land and were then subjected to erosion; (2) from areas of the original surface of the Earth that had not been covered over by the first sediments and hence could be attacked by erosion; and (3) from igneous masses that had invaded the rocks of either of the two foregoing categories.

This condition of affairs has continued to the present time: sediments have been laid down and converted to rocks; later they were elevated to form land and were exposed to erosion; they in turn have yielded by their erosion new sediments, and so on. Thus there has been a circulation of material, and changed conditions have affected the minerals at each stage. Only the most durable minerals, such as quartz, are likely to survive during such a cycle. The first stage in the cycle is the journey from the land to the sea; the return journey is made by the lifting of the newly formed sedimentary deposit above sea level. The silicate minerals, which form the bulk of the detrital sediments, have traveled to the sea by traction, by saltation, and in suspension. The carbonate minerals, on the other hand, have made the journey mainly in solution. This means that the detritus derived from the erosional destruction of sandstone, for example, is carried away mechanically, whereas limestone, consisting essentially of calcium carbonate, is carried away during erosion mostly by going into solution. Some limestones, however, are found to be made up of detrital fragments derived from older limestones. Such detrital limestones are of comparatively small volume. Notable examples are the limestone fanglomerates that occur in the Triassic of the Atlantic seaboard. We conclude that detrital limestones were formed under conditions in which the limestone in the source area was being eroded mechanically faster than it could be dissolved.

Sedimentary rocks derived by the destruction of older sedimentary rocks are "second-cycle" rocks. Third-cycle rocks are known to occur; and rocks whose material has probably gone through more than three cycles exist, so long and eventful has been the geologic history of our planet.

Minerals of sedimentary rocks. Sedimentary rocks, in view of their mode of origin as just sketched, are made up in variable proportions of constituents that fall into four classes.

A. Relict minerals, survivors of the attack of weathering on the source rock.
B. Minerals formed by weathering, *e.g.*, kaolinite, bleached biotite.
C. Biogenic products, fragments of fossils, et al.
D. Chemical constituents formed at place of deposition.

Quartz, feldspar, and rock fragments are important in the coarser-grained detrital rocks; quartz, kaolinite, and mica in the finer-grained ones. Calcite and dolomite are common in the nondetrital sedimentary rocks, both chemical and biogenic.

In the fine-grained rocks formed from silts and the aphanitic rocks formed from muds and clays, the particles are so minute that mineral composition is megascopically of little value in identification, compared with color, texture, structure, hardness, and other qualities.

Classification of sedimentary rocks. Sedimentary rocks are classified according to physical characters and composition. In this book the following classification is adopted.

I. DETRITAL ROCKS.
    Coarse detrital rocks: rudites.
        Conglomerates.
        Breccias.
            Basal, or transgression breccias.
            Fanglomerates.
            Tillites.
    Medium-grained detrital rocks: arenites.
        Sandstone.
        Arkose.
        Graywacke.
        Quartzite (orthoquartzite).
    Fine-grained detrital rocks: lutites.
        Siltstone.
        Shale.
        Mudstone and claystone.
        Argillite.
        Loess.

II. NONDETRITAL ROCKS.
    Precipitates.
        Chemical precipitates: products formed by addition of a precipitant to the solvent.
        Evaporites: products of evaporation.
        Duricrust rocks.
    Organic.
        Zoogenic: made up of hard parts of animals, *e.g.*, crinoidal limestone.
        Phytogenic: made up of plant remains, *e.g.*, algal limestone.

Classification of sedimentary rocks is at present in an unsatisfactory state, chiefly because of the heterogeneity of these rocks. Shales grade into limestones on the one hand and into sandstones on the other, and no sharply defined boundary can be drawn between them. Many such gradations can be cited. It is also difficult to know just where to assign many rocks of mixed origins. Some limestones, for example, are of detrital origin, some are of chemical origin, and some of organic origin; many limestones are made up of material of all three modes of origin. The classification must be considered as based on clear and unmistakable types, which serve as center points around which allied rocks can be grouped.

Chemical composition of sedimentary rocks. The chemical and mineral composition of sedimentary rocks are not so clearly governed by definite laws as those of the igneous rocks. The rules governing the occurrence of certain minerals together in the sedimentary rocks are more complex, since the minerals have been brought together partly by chance, controlled or modified, however, by the factors of shape, grain size, and specific gravity. Chemical analyses of some of the more important sedimentary rocks are given in the succeeding descriptive chapter, because such analyses are useful in several ways, scientifically and technically.

CHAPTER X

DESCRIPTION OF DETRITAL SEDIMENTARY ROCKS

Sedimentary rocks of detrital origin consist of the mechanical rock waste—the detritus—that was derived from the destruction of older rocks, was transported by the various erosional agents, and was eventually laid down as sediments. Subsequently the sediments were converted into rocks by the deposition of a cement between the detrital particles or by the development of a bonding material. As already indicated, sedimentary rocks are divided on the basis of the size of their detrital particles into three classes (1) coarse-grained *rudites*, in which the grain size exceeds 2 millimeters in diameter; (2) medium-grained *arenites*, in which the grain size is between 2 millimeters and 1/16 millimeter; and (3) fine-grained *lutites* (from the Latin *lutum*, mud), in which the grain size is less than 1/16 millimeter. These distinctions based on grade size, though none too practical in megascopic work, are nowadays generally accepted.

COARSE DETRITAL ROCKS: RUDITES

Conglomerates. Conglomerates consist of rounded detrital fragments of rocks and minerals held together by a cement of finer material. (See Plate 28.) The rounded fragments range from the size of a pea up to that of large boulders. Accordingly, we have *pebble conglomerates*, in which stones ranging from 2 to 64 millimeters predominate; *cobble conglomerates*, in which stones from 64 to 256 millimeters predominate; and *boulder conglomerates*, in which stones exceeding 256 millimeters (10 inches) predominate.

The stones became rounded by impact, grinding, and abrasion during transport by water. They may consist of rocks of any kind, though the harder and more resistant varieties generally predominate, or they may consist of a single mineral, generally quartz or feldspar. The stones may be all of one kind or a mixture of several kinds of rocks or minerals.

The cement of conglomerates may also be of various kinds; it may be composed chiefly of consolidated sand, either purely siliceous or of mixed substances; it may be calcareous; it may be chiefly composed of clay; or it may consist of these substances mingled with iron oxide.

There may be a sharp contrast between the relatively large pebbles and the fine matrix in which they are enclosed; and if this contrast is pronounced and the matrix is abundant, the conglomerate is called *puddingstone*. On the other hand, there are gradations in size from the pebbles down to the constituents of the matrix. There is, of course, great variety in the color of conglomerates. In some the pebbles are sharply defined from the matrix by their colors. In others pebbles and matrix have the same general color.

In the older formations, especially those in strongly folded mountain regions where the strata have been greatly sheared, the pebbles of conglomerates are generally distorted and flattened into lenses or have been drawn out into spindle-shaped forms. This deformation is generally accompanied by mineralogical changes which may be especially noticeable in the cement. These changes mark the first stage in the conversion of conglomerates into gneisses and schists by metamorphism, as described in the following chapter.

Breccias. In a breccia the fragments that correspond to those in a conglomerate are angular in shape instead of round. (See Plate 28.) This lack of rounding indicates, if the material was transported by water, that the fragments have not been transported far from their place of origin. In other respects, what was said in regard to conglomerates applies also to breccias.

Among the commoner types of breccias are (1) *basal* (or transgression) *breccias*; (2) *fanglomerates*; and (3) *tillites*.

BASAL (OR TRANSGRESSION) BRECCIAS. Basal (or transgression) breccias are formed, for example, by the sea advancing over a cherty limestone region, whose weathered surface is littered with angular fragments of chert.

FANGLOMERATES. Alluvial breccias cemented into rock are called *fanglomerates*. They are made up of angular detritus, showing little or no sign of water wear. The detritus is devoid of sizing, being a more or less chaotic assemblage of fine and coarse material, including enormous blocks; it is unshingled and in "pell-mell" orientation, the axes of the fragments being tipped in all directions. The deposit as a whole is characterized by imperfect and nonpersistent bedding. These features indicate that a sedimentary breccia of this kind was laid down as a piedmont alluvial fan and was subsequently indurated.

ORIGIN OF CONGLOMERATES AND BRECCIAS. Conglomerates and breccias that are composed of rock fragments of a single kind are generally called by the name of that constituent. We thus have quartzite conglomerates and breccias, limestone conglomerates and breccias, etc. Volcanic breccias produced by the accumulation of fragments blown out during eruptive activity can be considered to be igneous rocks and have already been described (see page 224). The material may fall into water, however, and become rounded, assorted, and stratified, giving rise to *volcanic conglomerates;* such rocks are difficult, if not impossible, to distinguish from conglomerates formed by the erosion of land areas underlain largely by lavas.

Breccias are produced also as the result of the breakage and grinding of rock masses along faults on which powerful movement is occurring. The fragments thus formed may afterwards be cemented together into firm rock by material deposited from solutions circulating in the zone of crushed and broken rock. Such breccias are called *friction breccias*, or *fault breccias*, and of course they show no evidence of stratification.

Conglomerates are normally formed from gravels laid down by moving currents of water, which tend to carry away the lighter and finer material in suspension. Hence they represent the deposits of rapid rivers and estuarine currents. Also, when the sea advances upon a subsiding land mass, a beach deposit forms as the initial stage to later deposits. The waves throw the coarser material, the gravel, toward the upper part of the beach, and as the beach advances inland conglomerate is the first deposit laid down on the new sea bottom. Thus a conglomerate called a *basal conglomerate* or a coarse sandstone generally constitutes the first member of a new series of stratified rocks that rests unconformably upon a lower series. Conglomerates are common rocks and are everywhere distributed in the sedimentary formations.

Uses. On account of their coarse and irregular appearance and inhomogeneity, conglomerates have been little used for structural purposes, except in the roughest stone work, as in foundations and piers. Some breccias, which are compact and capable of a good polish, have been cut as ornamental stones, as, for example, a reddish conglomerate breccia from South Dakota.

Tillite. A tillite is a till that has become converted into rock. It is generally unstratified and consists of a chaotic assemblage of angular, subangular, and rounded fragments that range in size from the finest pulverized mineral fragments (mineral flour) to huge boulders. Generally the matrix in which the rock fragments and boulders are enclosed predominates in volume. Tillite is likely to resemble a cemented alluvial breccia, or fanglomerate. In fact, tillite can be distinguished from fanglomerate only by finding in it faceted and striated stones of glacial origin, or by finding that it rests on a glaciated floor.

Tillites have been recognized to occur in many of the geologic formations, from Precambrian time onward.

Some notable tillites are the Early or Middle Precambrian *Gowganda tillite* of Ontario and Quebec and the Permian *Dwyka tillite* of South Africa, whose glacial features are as spectacular as those of the most spectacular modern glacial deposit.

MEDIUM-GRAINED DETRITAL ROCKS: ARENITES

Sandstones. Sandstones are composed of grains of quartz held together by a cement. (See Plate 27.) The grains range in diameter from 2 millimeters down to that of fine seeds; as they become finer the rocks grade into siltstones, just as on the other hand they grade upward into conglomerates. Consequently no sharp line can be drawn between siltstone and sandstone, and between sandstone and conglomerate. Some sandstones are very pure, consisting of quartz grains alone, but others contain much feldspar, generally orthoclase and microcline. They may contain also detrital garnet, magnetite, tourmaline, flakes of mica, and other minerals. Generally, as can be seen with the lens, the grains tend to be angular, but in some sandstones they are perfectly rounded. The general appearance of many sandstones as determined by their granular texture is much like that of cube sugar. As described under quartzite, the fracture in a sandstone takes place chiefly in the cement, causing the grains to stand in relief; this characteristic gives sandstone its sugary appearance and sand-papery feeling.

Sandstones differ much as to the cement that holds the grains together, and thus different varieties are produced. In some the cement is silica (quartz or chalcedony); in some it is a carbonate—commonly calcite, but also dolomite or siderite; in some it is extremely fine argillaceous material or clay; and in others it is iron oxides, either reddish (hematite) or yellowish (limonite).

The colors are many: white, gray, buff to dark yellow, and brick red to reddish brown and brown are common; green, purple, and black are rare. The color depends largely on the kind of cement: in yellow, red, and brown sandstones, iron oxides predominate; in the lighter colored sandstones the cement is likely to be calcareous or argillaceous. Calcareous sandstones are readily detected because they effervesce when touched with acid, whereas argillaceous sandstones give the characteristic odor of clay when breathed upon. Green color is due to glauconite or, more rarely, to admixed chlorite. Some sandstones are almost devoid of a cement.

Most sandstones are porous and permeable. Their porosity depends largely on the amount and character of the interstitial cement. Thus the ratio of the volume of pore space to the volume of the rock, *i.e.*, the porosity, ranges from 5 to 30 per cent or more. Since the individual pore spaces are relatively large, sandstones are permeable, allowing fluids—gas, oil, and water—to move through them. Because of their porosity and permeability sandstones are the common reservoir rocks from which oil is obtained in oil fields.

The same characters also determine to a large degree the other physical properties and explain their variations: thus the weight per cubic foot ranges from 125 to 175 pounds and the crushing strength from 1500 to 15,000 pounds per square inch. The bulk specific gravity ranges from about 2.0 in highly porous sandstones to about 2.7.

The chemical composition of sandstone varies considerably. The chief constituent is silica; the proportions of the others depend on the nature of the associated minerals and cement. Analyses of some prominent sandstones used for building purposes are shown in Table 1.

TABLE 1

ANALYSES OF SANDSTONES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	CO_2 etc.	Total
I	99.4	0.3		...	...	...	...	...	0.2	...	99.9
II	86.6	8.4	1.6	...	...	trace	0.7	2.4	...	...	99.7
III	92.9	3.8	trace	0.9	trace	0.3	0.3	0.6	1.2	...	100.0
IV	69.9	13.2	2.5		trace	3.1	5.4	3.3	1.0	...	98.4
V	87.1	3.9	1.3	...	1.1	2.7	0.8	1.3	0.5	1.4	100.1
VI	90.7	4.6	0.4	0.1	0.1	0.1	2.8	0.5	0.4	0.3	100.0

I. White, very pure Potsdam sandstone, Ablemans, Sauk Co., Wisconsin.
II. Lake Superior brownstone, Houghton, Bayfield Co., Wisconsin.
III. Sandstone, light gray, Berea, Ohio.
IV. Brownstone (arkose), Triassic, Portland, Connecticut.
V. Sandstone, Triassic, near Liverpool, England.
VI. Bunter sandstone, Heidelberg, Germany.

The content of alkalies points to the presence of feldspar (or mica) in the sandstone; in IV the amount of feldspar thus indicated is large.

The structure of sandstones is essentially that of the stratified rocks. Some are thinly laminated (Plate 27); others are thick bedded, and individual beds may show a very uniform texture and be practically free from any evidence of stratification. Sandstones of the latter type are valuable for structural purposes because of their homogeneity and capability of being dressed equally easily in all directions; they are often called *freestones.*

Sandstones are frequently distinguished according to the nature of their cement or of the admixed material. Thus there are *calcareous* sandstones, *argillaceous* sandstones, *ferruginous* sandstones, and *siliceous* sandstones. *Micaceous* sandstones contain considerable muscovite, the flakes of which are parallel to the bedding and induce a more or less ready cleavage in the rock; fracture surfaces are generally somewhat silvery in appearance because of the mica films coating them. *Grit* is a coarse-grained sandstone in which the particles are more or less sharply angular and the cementing material is, as a rule, quite siliceous. In some siliceous sandstones silica has been precipitated upon the rounded or angular quartz grains in crystalline position, thus converting them outwardly into crystals; examination with the lens reveals the crystal forms and faces of the little "regenerated quartzes"; such sandstones are known as *crystal* sandstones.

Green sandstone is a variety full of grains of glauconite, which give it a general greenish color. Some of these rocks are friable, indeed scarcely coherent, as in the Cretaceous formations of the Atlantic border, especially in New Jersey. They are then called *greensand*. They commonly contain, besides the sand and glauconite, iron oxides and fossil shells, either whole or fragmentary. These greensands have been used as fertilizers to supply potash to depleted soils. Analyses of typical greensands from New Jersey are given in Table 2.

TABLE 2

ANALYSES OF GREENSANDS

SiO_2	P_2O_5	SO_3	Al_2O_3	$Fe_2O_3 + FeO$	MgO	CaO	K_2O	H_2O	Total
34.5	1.2	1.3	6.0	31.5	2.2	2.5	1.5	18.8	99.5
51.2	0.2	0.4	8.2	23.1	2.0	0.5	7.1	6.7	99.4

USES OF SANDSTONE. Sandstone is widely used for constructional purposes. The ease with which it is worked and the large size of the blocks that can be quarried make it particularly valuable for this purpose. Thus in the United States many of the buildings of the eastern cities are wholly or in part built of red-brown sandstone, generally called "brownstone," obtained from the Triassic areas of the Atlantic border. The city of Edinburgh in Scotland is largely built of the Carboniferous sandstones of that region. Because of the insoluble nature of the iron oxide forming their cement, red and brown sandstones retain the details of fine cutting and carving for architectural effects much better in moist climates than do the lighter-colored gray or buff sandstones. The latter are likely to have a calcareous cement, which dissolves under atmospheric attack, causing the stone to crumble, and thus in the course of years the fine details of carving are spoiled. Many examples of this can be seen in the older cities where expensive and beautiful buildings have been much injured. If possible, in building, sandstone should be laid with its stratification in a position like the one it had in the quarry bed, as it is then much less likely to flake or spall.

ARKOSE. Arkose is a special variety of feldspar-bearing sandstone that simulates a granite in appearance. If mica also is present, the resemblance to granite is heightened. The constituent particles of quartz and feldspar are usually sharply angular. The feldspar is generally microcline and orthoclase, but plagioclase arkoses are known. Under a lens the irregular, angular shape of the particles distinguishes arkose from a granite. The mineral composition and the angular shape of the grains show that the material was derived from granite that was disintegrating mechanically faster than it was decaying chemically and that it was transported but a short distance before being deposited. Arkoses grade into conglomerates and breccias if some of the particles are somewhat bigger. They occur in many of the different geologic formations.

GRAYWACKE. Graywacke is a sandstone-like rock, generally gray and less commonly greenish or black, which, in addition to quartz and feldspar, contains rounded or angular bits of other rocks, such as fragments of shale, slate, chert, quartzite, granite, felsite and basalt. It is in reality a strongly indurated, impure sandstone. By increase in size of some of the component particles it grades into fine-grained conglomerate. The cement, as in sandstones, is generally small in amount and is generally argillaceous, but in some graywackes it is siliceous or calcareous. Graywackes, if fine grained and compact and largely composed of feldspathic material, may be difficult to distinguish from some felsites in the hand specimen, but close examination with a good lens generally shows their detrital nature.

Graywacke occurs in the Mariposa formation of the Sierra Nevada and in the Franciscan formation of the California Coast Ranges. It forms a belt 1000 miles long that extends along the Alaskan coast from Sitka to Cook Inlet. In the Sitka region the graywacke forms massive beds hundreds of feet thick and entirely devoid of stratification, so that it closely resembles an igneous rock. In Ontario and Quebec graywacke is common in many formations of Precambrian age.

ORTHOQUARTZITE. Sandstones that are so firmly cemented that they fracture across the grains instead of around them as normal sandstones do are commonly called quartzites. To distinguish them from quartzites of metamorphic origin (*metaquartzites*) they are called *orthoquartzites.*

FINE-GRAINED DETRITAL ROCKS: LUTITES

Siltstone. Siltstone is silt that has been converted into rock. It is intermediate in grain size between sandstone and shale. The coarser shales, in which minute quartz grains are abundant enough to give the rock an appreciably gritty feel, can be classified megascopically as siltstones.

Shale. Shale is a soft, minutely grained argillaceous rock, which has a more or less well-defined fissile structure. The characteristic fissility, which causes the rock to split into flat, shell-like fragments, is parallel to the bedding and is determined by the more or less parallel arrangement that the constituent flaky minerals assumed during sedimentation, an arrangement that was intensified by the subsequent sedimentary loading. During the compaction caused by the weight of the overlying sedimentary beds, the flakelets were rotated into positions more closely parallel to the bedding. Shales that were subjected to folding and pressure have assumed a slaty cleavage. This superinduced structure does not necessarily coincide with the stratification but may be inclined at any angle to it. Such transformed shales are slates or phyllites; they are described among the metamorphic rocks. The distinction between rocks having slaty cleavage and shales should be clearly noted, as shales and slates are often confused.

Shales are too fine grained for the component particles to be determined with the eye or the lens, or even with the microscope. By X-ray and chemical analysis they are found to be made up mostly of "clay minerals" (kaolinite, montmorillonite, and related substances), with which may be associated much white mica, sericite, and bleached biotite. These flaky minerals are generally accompanied by tiny particles of quartz and other minerals. As the amount of quartz increases and also the grain size, shales pass over into siltstones, as has already been mentioned. There is also a complete transitional series between clays and shales, in which the relative firmness and the fissility increase in the shale end members.

Clay is the material from which shales are derived; it is not a rock as that term is used in this book, but on account of its close relationship to shale it is briefly described here.

Clay when dry is a fine, earthy, lusterless substance, giving a characteristic "argillaceous" odor when breathed upon. It clings to the tongue. When strongly rubbed to a powder between the fingers, it produces a soft, greasy, lubricated feeling, thus differing from loess and similar-appearing materials. It absorbs water readily and becomes plastic. Shales have lost this power to become plastic. If pure, clay is white, but it is generally colored red or yellow by iron oxides, forming red and yellow *ochers*, or it is colored gray, blue, or black by an organic substance. Colors may be evenly distributed or irregularly blotched through the clay.

Shales are soft; they can be cut more or less readily with the knife and are brittle and crumbly. Like clays, they exhibit many colors: white to buff or yellow, red to brown, purple, greenish, and gray to black; and these colors are determined by the same causes as in clays. Grays are perhaps most common. Clays contain accessory minerals, such as carbonates, gypsum, rock salt, and pyrite. Some of these substances occur in the form of concretions, which may attain large size, up to several feet in diameter.

The chemical composition of shales is somewhat varied, depending on the relative proportions of clay minerals, relict minerals, and authigenic constituents. The analyses in Table 3 serve to show the general chemical character. The significant chemical features are high alumina content relative to lime and alkalies, iron largely in the ferric state, and high water content—all direct consequences of the fact that shales are formed of material derived from the weathering of preexisting rocks, involving oxidation, hydration, and the formation of aluminum silicates.

TABLE 3

ANALYSES OF SHALES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
I	55.0	21.0	5.0	1.5	2.3	1.6	0.8	3.2	8.1	1.9	100.4
II	60.6	16.4	4.9	...	1.4	1.6	0.9	3.0	9.7	1.5	100.0
III	61.2	15.6	1.4	3.0	4.2	3.4	0.4	6.7	2.7	1.1	99.7
IV	53.6	17.6	4.1	3.7	5.2	2.3	2.5	2.2	8.5	0.2	99.9

[1] XyO represents carbonaceous matter, CO_2, and small amounts of other substances.

I. Cambrian shale, Coosa Valley, Cherokee Co., Alabama.
II. Cretaceous shale, near Pueblo, Colorado.
III. Devonian shale, Morenci district, Arizona.
IV. Cretaceous shale, Mount Diablo, California.

There are many varieties of shales, depending chiefly on the presence of accessory materials. Some contain much organic matter, mostly carbon, and are called *carbonaceous shales*. They are black and, by increase of carbon, grade into coaly shales, shaly coals, and so on into coal. They are a common type and occur associated with coal and also independently of it; in places they extend over wide areas and are of great thickness.

Shales containing carbonaceous matter that yields oil on destructive distillation are called *oil shales*. Although called oil shales they contain no oil as such, but they yield oil on being heated in retorts at a high temperature. They are now attracting much attention and have enormous potentialities as sources of gasoline and petroleum. The most notable example is the Green River shale, an Eocene lake deposit, occurring in Colorado, Utah, and Wyoming; it constitutes a potential reserve of 100 billion barrels of oil, more than three times the total amount of oil so far produced by the oil fields of the United States.

Other varieties of shales contain carbonates, especially calcite; they are known as *calcareous shales*. By increase of calcareous content they pass into shaly limestones. They are commonly associated with limestones. These calcareous varieties are detected by their ready effervescence with acids. *Alum shale* is a variety full of pyrite or of sulfates resulting from its alteration; it has been used for the manufacture of alum.

Bentonite is a remarkable clay rock, recognizable by its extraordinary power of swelling from three to ten times its original volume when immersed in water. It is white in color and soft and unctuous to the feel, and it may resemble talc in appearance. It is largely composed of the mineral montmorillonite, a highly hydrated silicate of aluminum of the ideal formula $Al_2(OH)_4O_2Si_4O_6(OH)_2 \cdot nH_2O$, but in nature the aluminum is partly replaced by Mg and Fe. Bentonite commonly occurs as interbeds in marine shales, having formed from the alteration of tuffs made up of glass shards of moderately siliceous composition (60 to 68 per cent of silica).

USES OF CLAY AND SHALE. The use of clay in making bricks, tiles, and pottery is well known and needs no comment. Shale has no value for structural purposes. Along with clay, it is of great value, and is used in large amounts in many places, as a material in the manufacture of Portland cement by being mixed with the proper proportion of limestone and calcined. Shale or clay of the general composition shown in analysis II in Table 3 is the kind most suitable for this purpose.

Clays and shales are abundant in all parts of the world where unmetamorphosed stratified formations are found, so that their occurrence needs no special description.

Mudstones and claystones. Sediments consisting of extremely minute mineral particles that have become compacted to rock without taking on a fissile structure are called *mudstones*. Most rocks of this kind are indurated clays and can appropriately be called clay mudstones, or *claystones*. Indurated calcareous muds are termed calcite mudstones, or, perhaps more elegantly, calcite lutites, or *calcilutites*. Chamosite mudstones, as explained on page 279, constitute important sedimentary iron ores in England.

Argillites. Hard, indurated shales devoid of fissility are called *argillites*. They are like slates but do not have slaty cleavage. Their induration is probably the result of the dehydration and crystallization of the mineral gels originally present in minor amount. Many argillites are distinguishable only with difficulty from cherts and felsites.

Loess. Loess is a material of pale- to buff-yellow color, running into brown, and of exceedingly fine grain, averaging 0.05 millimeter. It is friable and if coherent has about the consistency of ordinary chalk; it has rather a harsh feel when rubbed between the fingers. It is remarkably homogeneous in appearance and is devoid of stratification. Loess consists chiefly of angular grains of quartz, mixed with minute particles of other minerals, and a calcareous cement, the amount of which may attain as much as 30 per cent. The carbonate produces a brisk effervescence in acid, which quickly ends. Characteristic also are high porosity (greater than 40 per cent) and high permeability, which cause the loess to absorb water freely. A loess from Kansas City, Missouri, has the following chemical composition:

SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	CO_2	XyO[1]	Total
74.5	12.3	3.3	0.1	1.1	1.7	1.4	1.8	2.7	0.5	0.4	99.8

[1] XyO represents minute quantities of P_2O_5, TiO_2, SO_3, MnO, and C.

Loess is widespread in the Mississippi Valley: in the States of Ohio, Indiana, Illinois, Iowa, Kansas, Nebraska, Arkansas, Missouri, Tennessee, Kentucky, Alabama, Louisiana, Mississippi, and Oklahoma. It occurs also in Europe, especially in the valleys of the Rhine and its tributaries, lying in isolated patches on the upper hill and mountain slopes, and in the same way in the Carpathians. It covers an enormous area in northern central China, with thicknesses attaining several hundred feet. The yellow color that the eroded loess imparts to the Hoang-ho (Yellow River) and to the Yellow Sea, into which the river discharges has given them their names.

Loess is now generally accepted as an eolian, that is, a wind-laid deposit of dust. This origin is shown by the lack of stratification in the loess; by the manner in which it is spread over the landscape, covering former hills and valleys; by the remains of land animals it contains, especially the shells of land snails; and by the small, vertical tubes encrusted with calcite running through it, caused by the roots and stems of former vegetation—grasses and mosses. In places, however, where it fell into or was washed down into former lakes, ponds, and streams, such reworked loess is stratified. The material of the loess in America and Europe is supposed to represent the finely ground rock powder produced by the Pleistocene ice invasion; conformably with this idea, four loess deposits have been found in places corresponding in ages with the four successive ice sheets. The Chinese loess, however, appears to have been a dust derived from the Gobi Desert.

A characteristic feature is the common occurrence of concretions of calcium carbonate and of iron oxide. They assume bizarre shapes like those of flint nodules in chalk. The perpendicular tubules are thought to give to loess the vertical cleavage that imparts to loess its most striking field character, its capacity to stand up in vertical cliffs. The name Council Bluffs is a reminder that the Missouri River thereabouts is flanked by high vertical bluffs, which are determined by a bed of loess 100 feet thick.

In summary, three factors were necessary to form a loess deposit: (1) a source of dust; (2) wind transport; and (3) a grassy or a moss-covered catchment area—steppe or tundra—to hold the dust in place after it was deposited. From the foregoing it is evident that loess is not a rock name. Loess converted into rock is loessite, but during this conversion the characteristic high porosity would disappear and the rock would be practically indistinguishable from a siltstone. Because of this difficulty, few rocks from the older formations have so far been interpreted as "dust-rocks."

CHAPTER XI

DESCRIPTION OF SEDIMENTARY ROCKS, MAINLY NONDETRITAL

CALCAREOUS ROCKS

LIMESTONES

General features. The chief calcareous rocks are limestone and dolomite. They have the property in common of being composed of the carbonates, calcite ($CaCO_3$), and dolomite [$CaMg(CO_3)_2$]. Those composed wholly or largely of calcium carbonate are called *limestones*; those composed wholly or largely of the mineral dolomite are called *dolomites.* Intermediate varieties are respectively called dolomitic limestones and calcitic dolomites. The calcium carbonate was primarily separated from water by being rendered insoluble by inorganic agents or by the action of living organisms, which either abstracted the calcium carbonate from the water in which they were living or precipitated it by means of chemical compounds produced by their metabolic processes.

Carbonate rocks are wholly soluble in hydrochloric acid if they consist entirely of carbonates. Impure varieties leave more or less of a residue, consisting chiefly of sand, clay, or silica. If dolomite is present the acid must be heated, unless the rock is finely pulverized. Carbonate rocks have a hardness of less than 4; hence they can readily be scratched or cut with the knife.

Limestones are the most abundant carbonate rocks. They range from aphanitic to coarsely phanerocrystalline. They are white, gray, dove color, bluish gray, dark gray, or black; red is less common. Yellows and browns result from iron oxide, grays and black from organic matter. Fine-grained limestones have an even fracture, and the most minutely grained have a conchoidal fracture.

Limestone effervesces briskly with hydrochloric acid, and feebly with vinegar (acetic acid) and lemon juice (citric acid). It is easily scratched with the knife and the less coherent varieties are friable to the finger nail. Its specific gravity ranges from less than 2.0 to 2.7, depending on the amount of impurities present and on the porosity of the rock. On exposed surfaces limestone, because of its solubility, tends to show the effects of solvent action: edges and corners are rounded off, and fissures and joints have been widened. It occurs in individual beds ranging from a fraction of an inch to more than 100 feet thick.

Some limestones consist wholly of grains of calcite; others have an interstitial claylike cement.

Table 1 gives analyses of limestones. I and II are analyses of the same limestone formation, showing the variation in composition; III is an impure variety containing considerable clay; IV is an average limestone; and V is ideally pure limestone.

TABLE 1

ANALYSES OF LIMESTONES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	H_2O+	XyO[1]	CO_2	Total
I	1.06	0.04	0.22	0.23	2.20	52.25	0.01	0.62	43.44	100.07
II	1.10	0.06	0.03	0.21	7.45	46.44	0.02	0.21	44.60	100.12
III	19.81	7.35	2.41	...	2.18	35.76	...	...	31.74	99.25
IV	5.2	0.8	0.5		7.9	42.6	0.6[2]	0.9	41.6	100.1
V	...	...	...	...	...	56.0	...	...	44.0	100.0

[1] XyO represents small quantities of organic matter and other constituents.

[2] Includes organic matter.

I. Columbus limestone, Sylvania, Ohio (analyst, D. Schaaf; sample represents the upper 20 feet of the formation).

II. Columbus limestone, Lakeside quarry, Lakeside, Ohio (analyst D. Schaaf; sample represents 12 feet of thin-bedded, highly fossiliferous limestone).

III. Argillaceous limestone used in the manufacture of natural rock cement, Hancock, Maryland.

IV. Average limestone (composite analysis of 345 limestones selected by G. K. Gilbert; analyst, H. N. Stokes).

V. Ideally pure calcium carbonate.

Limestones differ in strength according to their texture: firm, compact varieties are strong, and loose, porous ones are weak. Thus a fine-grained variety has a crushing strength of over 40,000 pounds per square inch, whereas others scarcely exceed 3000 pounds per square inch. The well-known gray oolitic limestone of Bedford, Indiana, has an average crushing strength of 4300 pounds. Any good, firm, and compact limestone has a strength far in excess of any load that it may be called on to carry in modern structures.

Limestones differ considerably in porosity; loosely cemented oolitic limestones are the most porous. The ratio of pore space to rock volume—the porosity—ranges from 25 per cent to practically nothing. The ratio between the weight of water a rock can absorb and its total porosity—the "effective porosity"—is in general much less than the total porosity, generally not more than one-half as much. The effective porosity is so much less because many of the pores are sealed off by cement and do not form parts of a freely intercommunicating system.

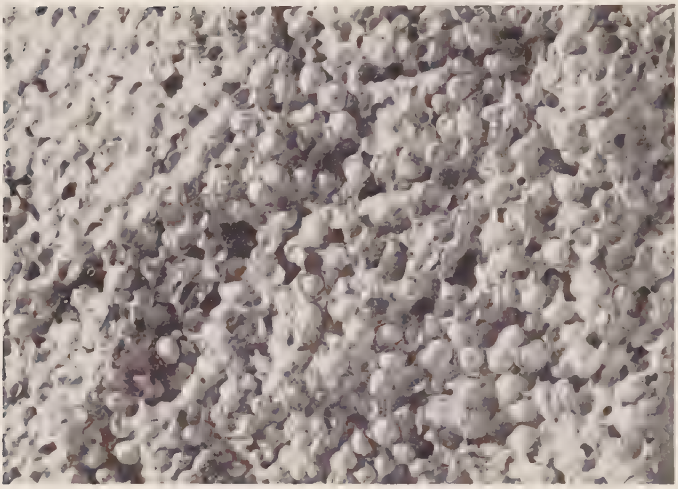
Limestones and other carbonate rocks are of common occurrence. Although the origin of most limestones is not ascertainable because at present we lack determinative criteria, the origin of some of the simpler kinds can be determined. Therefore we can define some varieties of limestone by their mode of origin. Organic limestones consist chiefly of the remains of calcareous fossils, which are likely to give them a distinctive character. The fossil remains comprise many kinds of organisms, among which are corals, crinoids, shells of mollusks, brachiopods, bryozoans, foraminifera, remains of sponges, etc. The fossils may be so abundant that the entire rock is composed of shells or of the hard parts of one particular kind of organism, with just enough fine calcium carbonate between them to act as a cement. Examples of this are the limestones composed wholly of brachiopod shells in the Niagara formation of Silurian age in western New York; they are called "shell limestones." Limestones composed of corals are other examples of "organic limestones."

On the other hand, there are varieties determined by the presence of some impurity that gives them a distinctive nature. Thus limestone containing much clay is called an *argillaceous limestone*, and one containing much quartz sand is an *arenaceous limestone*; such rocks are transitional to shales and sandstones. Others, which are dark colored and yield a strong, disagreeable, bituminous odor when struck or broken, are called *bituminous limestones*; they contain considerable organic matter. Some, called *glauconitic limestone*, contain more or less glauconite in green grains. *Lithographic* stone is an extremely fine-grained homogeneous limestone. The flesh-colored stone from Solnhofen, Bavaria, remarkable for the well-preserved fossils it contains, is especially used in lithography. It is a very pure limestone supposedly formed by evaporation and precipitation in a lagoon.

Many limestones contain chert in the form of nodules, irregular masses, and ramifying veinlets. Locally the chert is so abundant as to form definite bands or layers in the rock. On weathered surfaces of outcrops, the chert stands out prominently in relief, and the soils derived from cherty limestones are filled with innumerable fragments of chert.

During the weathering of limestones the calcium carbonate is carried away in solution and the insoluble impurities are left behind. The impurities thus become concentrated and form clays or loams

PLATE 29.



A. COARSE OÖLITE, BOHEMIA.

B. COQUINA, FLORIDA.

colored deep red or yellow by the oxidation of the iron minerals originally present. Cherty limestones as already mentioned give rise to stony soils charged with angular pieces of chert. Residual soils derived from limestones are commonly very fertile—the blue-grass country of Kentucky is a famous example—and the general prosperous condition of such regions has given rise to the saying "a limestone country is a rich country."

Oölite. Oölitic limestone. Oölite is a well-characteristized variety of limestone consisting of minute spheres and ellipsoids presenting much the aspect of fish roe; the name, which comes from the Greek, means eggstone. The spherules range in size from very minute—about the size of a pinhead—to as large as a pea. Those as large as peas are called pisolites. The larger spherules can generally be seen to have a concentric shelly structure and thus to consist of successive coats. A coarse oölite, or pisolitic limestone, from Bohemia is shown in Plate 29. More or less limy cement generally binds the oölites together.

The individual oölites as a rule contain some object, such as a bit of shell or a grain of sand, that served as nucleus around which the coats of carbonate were deposited. On the shores of Great Salt Lake at the present time oölitic sands are forming from the waters, which are charged with calcium and other dissolved salts. As the particles are rolled on the beach or agitated in the water, they become concentrically coated and assume spherical form. By a similar process oölitic grains are forming in springs charged with calcium salts, as at Carlsbad in Bohemia. The concretionary structure is best seen under the microscope; it is rarely sufficiently coarse to be seen by the unaided eye but can be sometimes made out with a lens.

Oölitic limestones constitute large and important formations, of great thickness and of different geologic ages. They are especially important in the Jurassic strata of England and elsewhere in Europe. Oölitic structure is prevalent also in many American limestones. Imperfectly cemented oölitic limestones are highly porous and can therefore do duty as reservoir rocks in oil fields. The highly productive Smackover Field in Arkansas draws its oil from such an oölitic limestone.

Chalk. Typical chalk is soft, white, and friable. A well-known use for it is making blackboard crayons. Most chalk is pure white, but some is gray, flesh color, or buff. It consists of a fine calcareous powder, which under the microscope is found to be made up chiefly of minute plates and discs of planktonic calcareous algae mingled with tiny shells of foraminifera. Present also, but in minor amount, are fragments of shells of larger organisms, as well as siliceous spicules of sponges, shells of diatoms and radiolarians, and sporadic minute fragments of various minerals. It is the siliceous material of the sponge spicules, diatoms, and radiolaria that has become concentrated into the nodules of flint so common in certain beds of chalk. Chalks, in spite of their fine grain, are very porous, some of them absorbing as much water as 20 per cent of their weight.

Chemically, chalks are pure calcium carbonate, as shown in Table 2.

TABLE 2

Analyses of Chalks

	$CaCO_3$	$MgCO_3$	SiO_2	$(Fe,Al)_2O_3$	H_2O	Total
I........	94.2	1.4	3.5	1.4	0.5	101.0
II........	96.4	1.4	1.6	0.4	0.2	100.0
III........	98.4	0.1	1.1	0.4	...	100.0

I. White chalk, White Cliffs, Little River, Arkansas.
II. Lower Cretaceous chalk, Burnet Co., Texas.
III. White chalk, Shoreham, Sussex, England.

Chalk was formerly thought to have been formed as a foraminiferal deposit on the bottom of the deep sea, because it resembles the calcareous oozes or muds now accumulating at great depth in the oceans. It has not been formed in this way, however, as it contains in places fossils indicative of shallow water. This and other facts demonstrate that chalk was formed by the accumulation of microscopic spheres (calcareous coatings of unicellular plants) in clear, warm, shallow seas into which no land-derived sediments were being delivered.

Chalk occurs extensively in Europe: in England, where it forms the "white cliffs of Albion"; in Germany; and in France. Its striking development in these countries as the product of a distinct geologic period caused that period to be named Cretaceous (Latin *creta*, chalk). Chalk is widely distributed also in Cretaceous formations in Nebraska, Alabama, Arkansas, and Texas.

Marl. Marls are loose, earthy materials consisting chiefly of calcium carbonate or dolomite intermingled with more or less clay. They are commonly gray, but many are yellow, green, blue, or black, and some have pronounced colors due to special substances, such as iron oxide and organic matter. They show all gradations into clays and shales. The carbonate content is readily detected by its effervescence in acid.

Varieties are named according to the special substances that they contain in addition to those just mentioned. Thus *sandy marl* is full of grains of quartz sand or of other minerals; *shell marl* is a whitish, earthy material made up of fragments of shells of organisms mingled with clay.

Greensand marl is described under sandstones.

The chemical composition of marls varies greatly. The following analysis of a compact marl from Colorado is given as an example of a typical marl.

$CaCO_3$	$MgCO_3$	SiO_2	Al_2O_3	Fe_2O_3	MgO	K_2O	H_2O	Org. Sub.	XyO[1]	Total
21.6	1.7	45.9	13.2	3.9	1.3	2.3	5.4	3.5	1.2	100.0

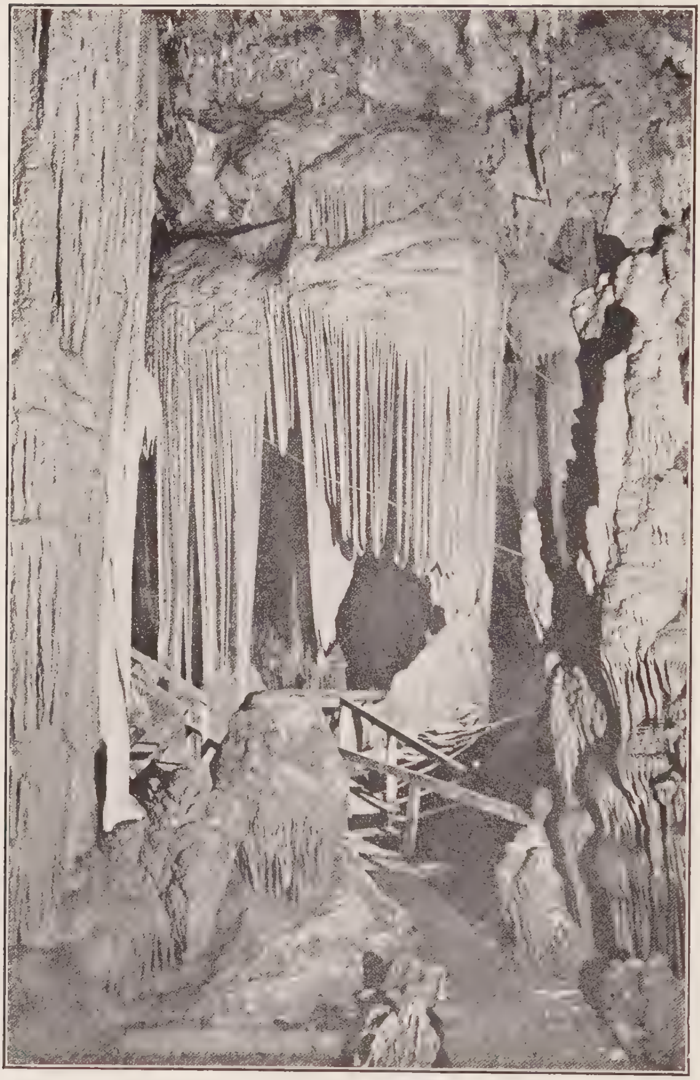
[1] XyO represents small quantities of TiO_2, Na_2O, and P_2O_5.

Coquina. A whitish, fragile rock now forming on the coasts of Florida consists of shells and their fragments of all sizes somewhat loosely packed and cemented together; it is known as *coquina*, from the Spanish word for shell. (See Plate 29.)

Travertine. It was shown earlier how the material of the land surface is in part made soluble during weathering, and having been made soluble, it is then dissolved and carried away to the sea. Solvent erosion of this kind is especially active with regard to calcium, which travels to the sea as sulfate and carbonate, the latter being much the more abundant. This calcium carbonate is derived not only from preexistent carbonate rocks but also from the calcium silicate minerals in igneous and metamorphic rocks, which under atmospheric attack are converted into carbonates.

Calcium carbonate is nearly insoluble in pure water. If, however, water contains carbon dioxide, calcium carbonate is converted into soluble bicarbonate. The amount of soluble bicarbonate formed and taken into solution depends on the amount of dissolved carbon dioxide. In regions where limestones or other carbonate rocks abound, the natural waters, under atmospheric pressure, attack such rocks and dissolve the calcium carbonate relatively slowly. In spring waters, especially thermal ones coming from depths, the pressure may be great and the amount of contained carbon dioxide large, and the quantity of dissolved calcium carbonate is proportionately large. Such waters lose most of the dissolved carbon dioxide on coming to the Earth's surface, and consequently the calcium carbonate in solution is deposited rapidly and copiously. From waters under atmospheric pressure the calcium carbonate is deposited by evaporation and therefore much more slowly. The deposition of calcium carbonate from warm waters is increased by the action of lowly forms of aquatic plant life (algae), which secrete calcium carbonate from the water.

PLATE 30.



STALACTITES OF TRAVERTINE IN LURAY CAVERN, VIRGINIA.

(U. S. Geological Survey.)

The rock formed by precipitation of calcium carbonate at the orifices of springs is called *travertine*, from the old Roman name of the town of Tivoli near Rome, where an extensive deposit of it occurs.

When calcium carbonate is deposited slowly, as by water dripping from the walls of caves, it is firm and compact and finely crystalline; this variety is called *dripstone*. Some is white, but most is tinted yellowish or brownish; it has a fibrous or concentric structure. The icicle-like deposits of dripstone formed by the downward growth from the roofs of caves are known as *stalactites* (see Plate 30). *Stalagmites* are the deposits that have grown upward where the down-dripping water has splashed on the floor of the cave. When the carbonate is deposited more rapidly, as by springs, the deposit is not perceptibly crystalline to the unaided eye and is porous to loose or earthy; when formed as a coating on vegetation it is cellular, bladed, or mosslike, as illustrated in Plate 31.

Many caves are celebrated for the number, size, and beauty of the stalactitic and stalagmitic deposits they contain (see Plate 30). Springs that deposit travertine are common the world over, especially in limestone regions. Warm carbonated waters issue chiefly in volcanic regions, however, or in those that have been volcanically active in the recent geologic past. The immense travertine deposits of Mammoth Hot Springs in Yellowstone National Park are justly famous (Plate 32), and others occur in California, Mexico, Italy, and New Zealand.

Mexican "onyx" or "onyx marble," which is used as an ornamental stone, is a travertine having a banded structure, beautifully brought out by its varied tinting through metallic oxides.

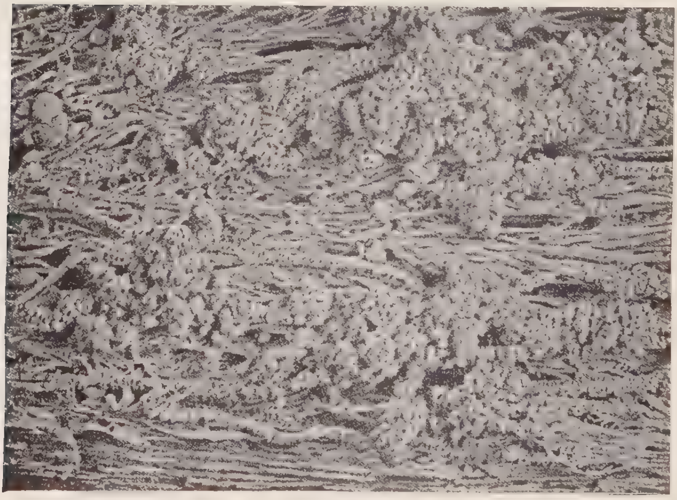
Uses of limestone. Limestone is used for structural purposes of all kinds. It is manufactured, by burning, into quicklime for mortar and cements. Large quantities also are used as a flux in smelting operations, as in the making of iron and steel. The use of argillaceous limestones containing 15 to 40 per cent of clay in the manufacture of natural hydraulic cements has become very important. Much larger amounts, however, are used in the manufacture of Portland cement, in which limestone of accurately known composition is mixed in carefully controlled proportions with clay or shale of accurately known composition, thereby assuring a product of uniform composition and properties.

DOLOMITES

Mineralogically, dolomite means a chemical compound of a definite composition, $CaMg(CO_3)_2$; geologically, dolomite is a name for a carbonate rock that consists dominantly of this compound.

PLATE 31.

A. TRAVERTINE, DEPOSITED ON VEGETATION, COLORADO.



B. TRAVERTINE, YELLOWSTONE NATIONAL PARK.
(U. S. Geological Survey.)

PLATE 32.

TRAVERTINE, MAMMOTH HOT SPRINGS, YELLOWSTONE NATIONAL PARK.

The description of the colors, texture, and other physical characters of limestone just given applies also to dolomite. In fact, mere inspection of a hand specimen is ordinarily not sufficient to determine whether a rock is a dolomite or a limestone.

Dolomite is somewhat harder than limestone. It is also somewhat heavier than limestone, its specific gravity being 2.87 instead of 2.71. If a carbonate rock is a pure dolomite, it effervesces in cold hydrochloric acid either not at all or only feebly; but if it is finely pulverized, it effervesces vigorously even in cold acid. The conclusive test is the determination of magnesium in the solution obtained by boiling the powdered rock in dilute hydrochloric acid.

The analyses in Table 3 show the chemical composition of some examples of dolomite. Analysis I is that of an extremely pure dolomite, as can be seen by comparing it with V, the theoretical composition of the mineral dolomite. The dolomite analyzed is further remarkable for the great thickness of such pure rock. In recent years such dolomites have been utilized as sources of metallic magnesium; ideally pure dolomite contains 13.18 per cent of magnesium. Analysis II is that of a dolomite containing a small amount of siliceous impurity. Analysis III represents a cherty calcitic dolomite, in which the silica is nondetrital. Analysis IV, in contrast, represents a dolomite containing abundant detrital grains of quartz and potassium feldspar (microcline and microperthite), and because of these impurities the dolomite has a low specific gravity—2.77.

TABLE 3

ANALYSES OF DOLOMITES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	CO_2	H_2O	XyO[1]	Total	Sp. Gr.
I.	0.2	0.0	0.0	0.1	21.8	30.4	trace	trace	47.7	0.0	0.1	100.3	....
II.	3.2	0.2	0.2	0.1	20.8	29.6	...	...	45.5	0.3	...	99.9	....
III.	9.2	0.3	0.1	0.1	13.6	34.4	0.0	0.0	41.8	0.1	0.3	99.9	....
IV.	25.5	2.2	0.6	0.4	14.8	21.6	0.9	1.3	32.0	0.4	...	99.7	2.77
V.	...	...	...	...	21.9	30.4	...	...	47.7	...	...	...	2.87

[1] XyO represents small quantities of other oxides.

I. Dolomite from Niagara dolomite, Luckey, Ohio (sample was taken across a thickness of 69 feet. Analyst, D. Schaaf).
II. Dolomite from Knox dolomite, Morrisville, Alabama.
III. Cherty calcitic dolomite, 10-foot members in Columbus formation, O'Shaugnessy Dam, Ohio (contains 62 per cent dolomite, 28 calcite, and 9 per cent quartz. Analyst, D. Schaaf).
IV. Arenaceous dolomite from Altyn formation, Alberta (contains 30 per cent of detrital grains of quartz and feldspar).
V. Ideally pure dolomite.

The origin of dolomites is a matter of much interest. It is clear that dolomites have been formed in two ways: (1) as syngenetic dolomite, which is a dolomite formed either by direct precipitation from sea water or by the alteration of limestone by reaction with sea water before the overlying stratum was laid down; and (2) as secondary dolomite, by alteration of limestone to dolomite long after the limestone was formed, a transformation called "epigenetic dolomitization." The most extensive dolomites have been produced by direct precipitation or by penecontemporaneous dolomitization. Proof of this mode of origin is seen in the fact that sharply defined beds of dolomite, which may be as little as a millimeter thick, are interbedded with pure limestone layers. If the dolomite had originated by the conversion of limestone beds into dolomite long after the beds had originally been laid down, the intervening beds of limestone would also have been changed to dolomite.

Dolomite of epigenetic origin generally occurs in smaller masses than syngenetic dolomite, but in some districts thick uniform masses of dolomite have been formed by the wholesale transformation of limestone. Whether this transformation, or dolomitization as it is called, has been syngenetic or epigenetic, the most convincing evidence for dolomitization is the occurrence in the dolomite of dolomitized remains of organisms that built their shells of calcium carbonate. A striking example of the often highly selective nature of dolomitization is vividly shown by the so-called limestone used in the interior decoration of the Parliament Building in Winnipeg, Manitoba. The gastropods and cephalopods in this rock have been partly or completely altered to dolomite, whereas the corals and brachiopods have remained immune. This stone is called "the tapestry limestone" because of the remarkable tapestry effect produced by dendritic mottlings formed by partial dolomitization of fragments of what was probably marine seaweed.

The mineral dolomite has a considerably higher specific gravity than calcite. During the conversion of a calcitic limestone to dolomite a reduction of volume, amounting to 13 per cent, should theoretically take place therefore. This reduction of volume helps to explain why some dolomites of replacement origin are porous rocks. Most dolomites, however, are not highly porous, so that the replacement either affected an aragonitic limestone, which would give no shrinkage, or took place volume for volume.

EVAPORITES

The more important deposits formed by evaporation of bodies of water are gypsum, anhydrite, rock salt, and calcium carbonate.

Gypsum, anhydrite, and rock salt are closely associated in mode of occurrence and origin. They have formed in bodies of sea water that became separated from the ocean by a barrier bar or other means and were under such an arid climate that the sea water became so highly concentrated by evaporation that its salts were precipitated. They are formed also in arid regions in lakes that are without outlets, in which evaporation consequently equals or exceeds the amount of inflow. As the water of all streams contains some dissolved salts, in land-locked lakes these salts accumulate until they reach their points of saturation and are precipitated.

Gypsum. "Gypsum" is used as the name both of a mineral ($CaSO_4 \cdot 2H_2O$) and of a rock. Gypsum rock is an interlocking aggregate of gypsum grains, ranging from microcrystalline to finely phanerocrystalline. As the result of partial recrystallization gypsum rock may take on a pseudoporphyritic fabric, whereby well-developed crystals of gypsum, measuring 20 by 30 millimeters on the clinopinacoid, occur scattered through an aphanitic matrix. Most gypsum rock is white, but some is yellow or red from pigmentary iron oxides, or dark gray from admixed clay or organic matter. It is soft and easily scratched with the fingernail. Other properties are given in the description of gypsum as a mineral.

The most common associates of gypsum are rock salt and anhydrite, because these rocks form under essentially similar conditions, as previously stated. Clay and bitumen are common impurities in gypsum. Varieties containing bituminous substances generally yield a disagreeable odor when broken.

It has been determined experimentally that gypsum is deposited from sea water when the temperature of evaporation is below 25° C, but anhydrite ($CaSO_4$) is deposited when evaporation takes place above 25° C. Gypsum and anhydrite can therefore serve roughly as geologic thermometers.

Gypsum and gypsum products are being used in industry in increasingly large tonnages. Gypsum is used in the manufacture of plaster of Paris; as partition block and other fireproof structural material; in large amounts as a minor ingredient in Portland cement to prevent it from setting too fast; and in the raw state as a fertilizer (land plaster). Handsome fine-grained white or tinted varieties are called *alabaster* and are cut into ornamental objects, such as vases.

OCCURRENCE. Gypsum is widely distributed in sedimentary rocks of various geologic ages as extensive beds, many of which are of great thickness. It is associated especially with limestones and shales. It commonly accompanies beds of rock salt and as a rule underlies the salt. It occurs also in other sedimentary formations, especially in clays and shales, and as lenticular masses or as isolated crystals, some of great size, as in the Cretaceous beds of the western United States.

Anhydrite. Anhydrite rock is a granular aggregate of anhydrite ranging in grain size from microcrystalline to phanerocrystalline. Some anhydrite rocks, in fact, are coarse enough so that the seemingly cubic cleavage of the individual grains can be seen. Anhydrites may be somewhat translucent and generally have a rather splintery fracture and shimmering or pearly luster. Most anhydrites are white, but as with gypsum, some are tinted reddish, yellowish, bluish, gray, or dark by iron oxides, commingled clay, or organic matter. Anhydrite is harder than gypsum but is easily cut with a knife. For other properties, see the description of the mineral anhydrite. The most commonly associated minerals are halite and gypsum, but locally it may contain many others—those stated under gypsum, for instance.

Anhydrite, like gypsum, which it frequently accompanies, occurs as beds in marine strata. It becomes hydrated on exposure to the air, changing into gypsum. Extensive deposits occur in Nova Scotia. Thick beds of anhydrite, as much as 175 feet thick, are common in Late Permian strata in west Texas. They occur also in Late Permian strata in Germany, where they attain thicknesses of 300 feet and, in places, 1600 feet.

The deposition of a bed of anhydrite 300 feet thick would require the evaporation of a column of sea water 660,000 feet high. The Permian sea was shallow, however, and therefore some such hypothesis as an "evaporating pan" continuously fed by influx of sea water must be invented to account for the unusual environmental conditions that allowed these thick masses of anhydrite to be deposited.

Rock salt. Rock salt is a crystalline aggregate of grains of common salt (halite, or sodium chloride). It ranges from fine to coarse grained. It is generally white, but some is red from inclosed minute lamellae of hematite, or gray from intermingled clay or organic matter. The properties of halite are mentioned in the chapter on rock minerals.

Associated minerals in rock salt are gypsum, anhydrite, and, rarely, quartz, carbonates, and pyrite.

Rock salt occurs in many parts of the world, in sedimentary rocks of all ages from the Cambrian onward. The beds differ greatly in thickness, from 1 foot to 400 feet or more. The great thickness of some salt beds, just like that of the anhydrite, cannot be explained as the result of the simple evaporation of an isolated body of sea water in an arid region. Subsidence must have been gradually going on; at first the less soluble gypsum or anhydrite would be deposited, and then the salt, leaving a "mother liquor," or brine, which contains the more soluble sulfates and chlorides of magnesium and potassium. If subsidence and lowering of the barrier should now occur, sea water could flow in from the open ocean and spread over the more concentrated sea water, whereas the heavier mother liquor would flow out below; thus the evaporating basin would be charged anew with sea water. If the barrier were again closed, for example, by waves building it up, as takes place along the coast of the Carolinas, the conditions would be repeated and additional deposits of gypsum and salt would be formed. Thus by repetitions of this process a great thickness of salt might be deposited. Finally, if the brine could not flow away, it also would evaporate and the potassium and magnesium salts dissolved in it would be deposited.

If a sea should evaporate without geologic disturbance, the salts would be deposited in the order of their solubility: (1) calcium carbonate ($CaCO_3$); (2) calcium sulfates—gypsum ($CaSO_4 \cdot 2H_2O$), if the temperature of evaporation was below 25° C, and anhydrite ($CaSO_4$) if it was above 25° C; (3) sodium chloride (NaCl); and (4) potassium and magnesium salts.

Rock salt occurs in extensive beds in New York, Michigan, Louisiana, Texas, Kansas, and other States. The Mid-Continent field (Kansas, Oklahoma, Texas, New Mexico) is estimated to contain 30,000 billion tons of salt. After long exploration, valuable deposits of potassium salts have been found in portions of this field. The vast deposits of rock salt in Germany are notable, having associated with them the world's greatest reserves of potassium salts. Interior drainage basins are present in all of the continents and contain salt lakes and deposits of salt. Great Salt Lake in Utah is our own famous example.

DURICRUST ROCKS

In many parts of the world in arid and semi-arid regions a surface layer of hard rock extends as a covering or carapace over large areas. These hard crusts have been termed *duricrusts*. In lateritic regions this crust may consist largely of iron hydroxides, and it has been expressively named by French investigators the "iron cuirass."

In limestone regions the duricrust is a fine-grained calcite rock resembling limestone; in America it is called *caliche*. The type locality in Tucson, Arizona. The basal portion of such caliche crusts is likely to consist largely of angular rock fragments held together by carbonate and consequently resembles concrete aggregate; hence this rock has been named *calcrete*. Palestine is notable for the thick caliche crusts that have formed there; the stall in Bethlehem in which Christ was born is a cave dug into the lower part, which is the soft part of the caliche crust.

Gypsum crusts have formed in Mesopotamia and in Egypt, up to 1 meter thick. In millions of holes this impure gypsum is dug out by the fellahs and taken by camel transport to the Nile where it is ground, and then part is used as fertilizer—"land plaster"—and the remainder is shipped to Cairo for use as gypsum mortar. The gypsum renews itself at the surface, so that after a lapse of years the abandoned pits can again be worked. Here we appear to have actual evidence of the generally accepted idea that duricrusts are formed by water rising to the Earth's surface, evaporating, and precipitating its dissolved salts.

In other regions have formed siliceous crusts called *surface quartzites*, and in South Africa certain facies of these surface quartzites are called *silcretes,* again by analogy with concrete. Vast areas of Australia are covered by a carapace of duricrust, and aluminous, ferruginous, calcareous, and siliceous duricrusts are recognized, the particular composition being evidently determined by the chemical nature of the underlying rocks.

SILICEOUS ROCKS OF BIOGENIC AND CHEMICAL ORIGIN

Silica is deposited from solution in water both by evaporation and by the activity of living organisms; possibly it is deposited also as a result of chemical reactions. The deposits thus formed, especially those of biogenic origin, are of great interest and importance. Because of their similar composition, the chemical and biogenic cherts are here included under one heading, but the group does not include the sandstones, which, though of siliceous composition, are of detrital origin. The material composing the siliceous rocks here grouped together occurs partly in the form of quartz (pure crystallized silica), partly as chalcedony (a microfibrous form of quartz), and partly as opal (noncrystalline silica containing more or less water).

Chert and flint. Chert is an aphanitic rock composed of cryptocrystalline silica or opal, or of both. The dominant form of silica in most cherts is microfibrous chalcedony. In color, cherts range from white, cream, or gray to jet black; yellow, red, and brown are less common. Obviously, color is of no help in identifying chert. The hardness (7), however, is distinctive; the fracture is conchoidal.

Chert occurs in masses that range from the smallest nodules to formations of great thickness and wide areal extent. When studied under the microscope, some cherts are found to contain the hard siliceous parts of organisms, chiefly those of Radiolaria, diatoms, and siliceous sponges. Cherts made up largely of the remains of Radiolaria are called *radiolarian cherts* or *radiolarites*. As many as a million Radiolaria may occur in a cubic inch of chert. A notable radiolarian chert is the Franciscan chert of the Coast Ranges of California and Oregon, which attains a maximum thickness of 900 feet.

Flint is a dark-gray or black rock, but despite its dark color flakes of flint are translucent on thin edges just like those of many felsites, which indeed flint may closely resemble. Flint consists of chalcedony and extremely fine-grained quartz. The chalcedony gives flint its well-known tenacity, or toughness. The coloring matter of flint is organic and disappears when a chip of flint is heated before the blowpipe. Flint can be considered as a special variety of chert. Chert is the term commonly used in geology for homogeneous siliceous rocks, regardless of color, that are of porcelanous, chalcedonic, or opaline appearance.

Flint occurs most typically in chalk as irregular nodules or concretions which range in size from that of a pea to extensive layers. The silica of the flint nodules was extracted originally from sea water by living organisms, chiefly diatoms, Radiolaria, and siliceous sponges, and the remains of these organisms were deposited in the accumulating calcareous mud. Later the disseminated remains of the organisms, consisting of opal, were dissolved by subsurface water containing carbon dioxide, which was moving through the chalk; the silica thus entrained, after traveling some distance, was precipitated in the gelatinous state around certain nuclei, where favorable conditions obtained; subsequently the gelatinous silica crystallized as chalcedony. The use of flint by aborigines for implements and weapons and for striking fire is well known.

Jasper. Jasper is the name sometimes given to red or reddish-brown chert. Some jaspers, for instance those of the Lake Superior iron districts, are interlaminated with layers and streaks of hematite. The intervening cherty layers are colored bright red by finely disseminated hematite. Jaspers of this kind constitute large rock masses, the so-called iron formations. The ferruginous chert of the Mesabi iron district of Minnesota is called *taconite*.

Novaculite. Novaculite is a remarkable variety of siliceous rock, which occurs in considerable beds in Arkansas. It is extremely fine grained, conchoidal in fracture, white or pale gray, semi-translucent, and wholly composed of silica. It is a radiolarian chert that has been metamorphosed to its present state by the folding to which it was subjected. It is used in the manufacture of hones and whetstones.

Geyserite, or siliceous sinter. Silica is commonly deposited around the orifices of hot springs and geysers in volcanic regions. The material thus deposited is known as *geyserite*, or *siliceous sinter*; it is a form of opal ($SiO_2 \cdot nH_2O$). If pure, it is white. How siliceous sinter is formed is well illustrated in the hot spring and geyser areas of Yellowstone National Park and New Zealand (see Plate 33).

Diatomite. Diatomite is a soft, white, chalklike, very light rock composed of innumerable microscopic shells of diatoms. Diatoms are exceedingly minute, one-celled aquatic organisms, which have the power, remarkable for a plant, of being able to swim about freely. They secrete a siliceous shell of great delicacy. Ten thousand species are recognized, comprising both fresh- and marine-water kinds. At times diatoms proliferate in incredible numbers, as veritable diatom epidemics, and their shells, accumulating at the bottom, give rise to extensive deposits.

Most varieties of diatomite are white, but some are pale yellow, brown, or gray. Diatomite is easily distinguished from chalk, which it may resemble, by its much lower specific gravity (about 0.4) and its noneffervescence with acid; it is distinguished from clay by its gritty feeling when rubbed between the fingers, and by its weak argillaceous odor or absence of odor. A more positive test is the effervescence produced when some of it mixed with sodium carbonate is fused before the blowpipe.

Loose, scarcely coherent material is called *diatomaceous earth*; the more coherent material is called *diatomite*. It is extensively used for insulating material, polishing purposes, and many other purposes. Beds of fresh-water diatomaceous earth of considerable magnitude occur in many places in the United States and elsewhere, generally as a layer in swamps that represent the fillings of former lakes.

Marine diatomites of Miocene age attain the great thickness of 5000 feet or more in the Coast Ranges of California. They are of additional interest because they are regarded as the source rocks from which much of the enormous amount of oil of that region was derived. In southern California the diatomites are soft, punky rocks, being in places quarried on a large scale for insulating material, but farther north, in the Coast Ranges of central California, they have been altered to hard cherts.

PHOSPHORITE: BEDDED PHOSPHATE ROCK

Phosphorite is a rock consisting essentially of tricalcium phosphate ($3CaO \cdot P_2O_5$). It resembles limestone, though it is somewhat heavier. It is generally so fine grained as to appear amorphous to the unaided

PLATE 33.

GEYSERITE, FORMING THE "CRATER" OF CASTLE GEYSER, YELLOWSTONE NATIONAL PARK. (C. D. Walcott.)

eye, but according to X-ray analysis it is made up of minerals that are crystalline, chiefly fluorapatite [$3Ca_3(PO_4)_2{\cdot}CaF$] and other allied calcium phosphates, such as dahllite, $3Ca_3(PO_4)_2{\cdot}CaCO_3$. It is commonly oölitic.

Phosphorite occurs as stratiform layers in marine beds, ranging in age from Cambrian to Recent. The principal domestic deposits occur in Idaho, of Permian age, and in Florida, of Cenozoic age. Phosphate rock is literally vitally necessary to the national welfare. Its chief use is as a plant food. When used as fertilizer it is first converted to superphosphate to increase its solubility and hence its availability to the growing plants. The raw materials for the manufacture of superphosphate are phosphate rock and sulfuric acid. The production of superphosphate has increased from 1,505,000 tons in 1900 to 8,000,000 tons in 1946. The domestic reserves of phosphate rock have enormously increased in recent years and now amount to 10 billion tons, enough for 3000 years at the present rate of consumption.

The origin of the bedded phosphate rocks is a problem of much interest. The phosphate rock commonly contains the remains of calcareous marine organisms, but these remains now consist of phosphate: manifestly, they have been replaced by phosphate. Was this replacement effected long after the deposition of the fossiliferous bed or did it take place concurrently with deposition on the sea floor? The phosphate beds in the Permian Phosphoria formation of Idaho and adjacent States had originally an extent of 175,000 square miles. This one important fact of itself practically proves that this particular phosphate bed is of sedimentary origin. Further light is shed on the problem of origin by the discovery that nodular phosphorite is abundant on the sea floor of the coast of southern California. This phosphorite appears to have been directly precipitated from the ocean water, but some foraminiferal and other calcareous shells have been replaced by phosphate, in complete analogy with what is seen in the Idaho phosphorite.

Calculations based on the known chemical composition of sea water indicate that sea water is essentially saturated with tricalcium phosphate; consequently, a small change in conditions might cause precipitation of the phosphate. Because of the notable fluorine content of marine phosphate rock, amounting to 2 or 3 per cent, addition of fluorine to the ocean during periods of heightened volcanic activity appears to be a possible explanation. Fluorapatite, which is the chief constituent of phosphorite, is more insoluble than tricalcium phosphate and would therefore be precipitated if fluorine were added to the ocean. During the height of its eruptive activity, Katmai in

Alaska exhaled nearly 200,000 tons of hydrofluoric acid a year. The suggestion has therefore been made that the voluminous precipitation of phosphate from the seas during the geologic past has been determined by excessive emission of hydrofluoric acid at times of vigorous volcanic activity. The precipitation of the fluorine-bearing phosphate has the biologic merit of removing from the sea the toxic element fluorine.

SEDIMENTARY IRON-ORE ROCKS

Beds of iron ore of marine sedimentary origin occur throughout the world in strata of many geologic ages. They constitute mineral resources of great economic value; in fact they make up nearly 80 per cent of the world's known reserves of iron ore. The ironstone, as the material of the iron-ore beds is called, may consist of one or more of the following iron minerals: hematite, goethite, chamosite (an iron chlorite), and siderite.

The Clinton iron ore, the basis of the iron and steel industry in the Birmingham district, Alabama, is a red ore made up chiefly of hematite. The hematite is partly in the form of oölites and partly as replacements of broken bits of fossils, mainly crinoids, corals, and brachiopods. Where oölites predominate, the ore is called oölitic ore; where the replaced fragments of fossils predominate, is is "fossil ore." The fossil ore is essentially a hematitized coquina. The iron ore occurs as one or more interbeds in the Clinton formation, which extends from Niagara Falls to Alabama. In places the ore beds are absent, but where present and 2 or 3 feet thick they are commercially valuable. Near Birmingham, Alabama, the main ore bed is 20 feet or more in thickness.

A stratiform mass of iron ore of this kind poses this genetic problem. The ore is clearly the result of replacement, but is it syngenetic or epigenetic? The conclusion that it is of syngenetic origin, *i.e.*, that the iron-ore bed had been formed before the hanging-wall bed was laid down over it, is based on four items of evidence: (1) absence of fissures in the underlying beds by which iron-bearing solutions could have gained access to the ore bed to effect its replacement; (2) the great areal extent of the ore bed and its constancy of composition—features foreign to deposits formed by replacement but consonant with sedimentary origin; (3) in places limestone occurs in sharp contact with the ore bed, an improbable relation if the ore bed itself was formed by epigenetic replacement of limestone; and (4) fragments of the ore itself occur in the hanging-wall bed—evidence that is of crucial importance if it can be found, because it proves alone that the ore bed is of syngenetic origin.

Another remarkable sedimentary iron ore is the Wabana iron ore of Newfoundland, one of the world's great iron fields. The ore is of Ordovician age, somewhat older therefore than the Clinton iron ore. The ore occurs as interbeds, sixty of them, distributed through a formation 500 feet thick and ranging from a fraction of an inch to 23 feet in thickness. It is generally oölitic, the oölites being very minute, ranging from $\frac{1}{10}$ to $\frac{1}{2}$ millimeter in diameter. The oölites, curiously enough, consist of alternating layers of hematite (ferric oxide) and chamosite (ferrous chlorite), which coat a nucleus consisting of a quartz grain or bit of shell fragment. Manifestly, the oxidizing power of the sea water was fluctuating while the iron minerals were being deposited. The ore beds show ripple marks and cross bedding. They contain marine fossils—trilobites and brachiopods. Finally, the item of crucial importance in determining the mode of origin of the ore is the finding of pebbles of oölitic iron ore in the shale bed that overlies the main ore layer. From this evidence here briefly outlined the conclusion is reached that the Wabana iron ore is of marine sedimentary origin and accumulated in a shallow sea, in which, despite the active precipitation of iron minerals, the concentration of iron salts was not high enough to be toxic.

The genetic conclusion reached as to the sedimentary origin of the Clinton and Wabana iron ores is of the highest practical importance. For it makes it possible to compute, with considerable assurance, the probable tonnage of ore reserves that exists underground far beyond the actual realm of observation. This great practical advantage accrues, of course, wherever any stratiform deposit can be shown to be of sedimentary origin rather than the result of epigenetic replacement. A further matter to be noted is that such genetic determinations are not based on hand specimens but require a penetrating study in the field.

Sedimentary iron ores comprise the chief iron resources of England. The most common type comprises chamositic mudstones. These mudstones, as their name implies, are extremely fine grained; they are greenish in color from the presence of the iron chlorite chamosite ($Fe_3Al_2Si_2O_{10} \cdot nH_2O$). The ores are made up of oölites of chamosite in a matrix of chamosite, clay, and siderite in variable proportions.

The sedimentary iron ores of England are so important nationally that they have been carefully studied.* They have been classified as

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* A. F. Hallimond, Iron ores: bedded ores of England and Wales. *Mem. Geol. Survey Great Britain,* 1925.

shown in Table 4, which is reproduced here in order to show in some measure the diversified nature of these rocks.

TABLE 4

CLASSIFICATION OF THE SEDIMENTARY IRON ORES OF ENGLAND

FERROUS IRON ORES

I. Chamosite mudstones and chamosite-siderite mudstones (oölitic or compact), *e.g.*, the Cleveland ironstone.
II. Siderite mudstone, nonchamositic (compact), *e.g.*, the coal measure ironstones.
III. Sideritic limestones.

FERRIC IRON ORES

IV. Ferric chamosite oölites (containing free oxides of iron).
V. Limonite oölites, *e.g.*, the Frodingham ironstone.
VI. Primary hematites, *e.g.*, the Crinoidal hematite of South Wales.
VII. Glauconitic rocks (not of commercial value as ores).

The seas, at various times in the geologic past, have manifestly been able to form extensive deposits of bedded iron ores. The concentration of iron in the sea water during these episodes, however, was not high enough to be toxic to plant or animal life. In the present ocean the concentration of iron is infinitesimally low—2 micrograms (0.002 milligram) per liter of sea water. The only iron compounds known to be forming from the seas at the present time are pyrite, which is being precipitated on the floor of the Black Sea, the deeper levels of which are without circulation, hence "poorly ventilated" and foul with hydrogen sulfide; and glauconite ($KFe^{+++}Si_2O_6 \cdot nH_2O$, where n is generally 3), which is found at the edges of the continental shelves, in cold-water currents (hence oxygenated) at temperatures between 3° and 15° C. In Monterey Bay, California, glauconite is being formed by the action of sea water on detrital biotite, which during its transformation swells to twenty times its original volume.

Another sedimentary iron-bearing mineral is greenalite ($9FeO \cdot Fe_2O_3 \cdot 8SiO_2 \cdot 8H_2O$), which resembles glauconite in appearance and composition, except that it contains no potassium and its iron is chiefly in the ferrous state. It occurs abundantly in the iron-bearing cherts of the Precambrian Biwabik formation in the Mesabi iron district, in Minnesota, which is by far our most productive iron-ore district. Greenalite occurs in greenalite chert in the Ordovician of southern Scotland, and it is reported to be an important constituent in Jurassic sandstones in southern Sweden. These recent finds of greenalite are interesting and important in showing that the mineral is not restricted to rocks of Precambrian age.

COAL AND OTHER CARBONACEOUS ROCKS

Interbedded with other stratified rocks of the geologic periods from Devonian down to the present are layers of carbonaceous matter, which, under the names of coal and lignite, represent the remains of vegetation that flourished where these beds now are. The formation of peat in modern lakes, swamps, and bogs and its occurrence in beds interstratified with sands and clays in recent delta deposits, as in the Mississippi delta, show us how these beds of coal were formed. Every step of the gradual transition can be traced: from the growing vegetation of today into peat, from peat into lignite, and so on into bituminous coal and into anthracite, and where affected by strong metamorphism into graphite, which is practically pure carbon.

The vegetal matter composing plants consists chiefly of carbon, hydrogen, and oxygen. Its decay is a process of oxidation caused by bacteria and fungi. The hydrogen is oxidized to water and the carbon to carbon dioxide, the oxygen of the air assisting that in the vegetal matter to effect the change. In this process most of the carbon is removed. However if the decay takes place under water access of the oxygen of the air is prevented and the process becomes anaerobic. Some of the carbon unites with some of the oxygen to form carbon dioxide; some of the hydrogen unites with the rest of the oxygen to form water; the rest of the hydrogen unites with some of the carbon to form marsh gas (methane); and the remainder of the carbon is left behind. This "intramolecular combustion," as it is called, is illustrated by the following equation in which the formula of cellulose, the dominant constituent of plant matter, is used.

$$\underset{\text{Cellulose}}{C_6H_{10}O_5} = \underset{\text{Carbon Dioxide}}{CO_2} + \underset{\text{Water}}{3H_2O} + \underset{\text{Methane}}{CH_4} + \underset{\text{Carbon}}{4C}$$

This change does not take place at once, or become complete, under water; it goes on gradually, and as the carbon dioxide and methane are evolved, the residual matter becomes richer in carbon and poorer in hydrogen and oxygen. The plant matter is thus converted into peat, and the peat is transformed by compaction under sedimentary loading and by further chemical change into lignite. The same process continues in coal beds, generating the deadly gases known to the miners as choke-damp (CO_2) and fire-damp (CH_4), and lignite thus changes to bituminous coal. Deep burial of the coal beds and folding of the strata, accompanied by heavy pressures and also the attendant rupturing and fissuring of the overlying beds, which permit the gases to escape, appear to favor the process of "coalification," and under such circumstances the coal is changed to anthracite, which is much higher in carbon than lignite, or even into graphitic coal, which is practically pure carbon. Thus the degree to which a coal has advanced from its most immature state—lignite—to the most mature coal—anthracite—depends on the geologic conditions to which it has been subjected.

Peat. Peat varies from the upper portion of the bed, where it is a brown to yellowish matted mass of interlaced fibrous material, strongly resembling compressed tobacco, in which the remains of plant leaves, stems, and roots are still recognizable, to the deeper, lower parts, where it is a dark-brown or black, compact, homogeneous mass, much resembling dark clay when wet. A dried, compact peat low in ash content from Germany has the following composition:

Carbon	Hydrogen	Oxygen	Nitrogen	Ash	Total
55.9	5.8	36.4	1.0	0.9	100.0

Under enormous pressure peat has been experimentally converted into a hard, black substance like coal. Peat is widely distributed and is used as a fuel where better fuels are not available. The impurities in peat are chiefly caused by the clay and sand washed into it while it was accumulating; even the purest peat has, like coal, a small percentage of ash that results from the mineral matter in the constituent plants.

Classification of Coals. Coals are classified according to *rank*, or degree of maturity, on the basis of their fixed-carbon content and physical properties. The differences in rank are due to the progessive metamorphism of vegetal matter from lignite to anthracite, marked by progressive reduction in the amounts of water and volatile matter and increase of fixed carbon in the coals of higher ranks. As there is a continuous gradation from the lowest rank to the highest rank, the division lines adopted are necessarily arbitrary. By "fixed carbon" is meant the carbon that remains after the moisture and volatile matter have been driven off from the coal.

Coals of the lowest rank comprise the *lignitic* group. They are highly immature coals. When taken from the mine they may appear to be perfectly dry, and yet they contain 30 to 40 per cent of water. On exposure to the atmosphere they lose most of this loosely held water; as a result they develop innumerable shrinkage cracks and crumble to pieces, and they become dangerously prone to take fire spontaneously.

The coals next in rank are the *subbituminous*. Their black color and nonwoody structure distinguishes them from the lignites, and their tendency to slack on weathering distinguishes them from coals of higher rank.

The coals of medium rank are the *bituminous*. They do not slack on weathering and generally have a prismatic jointing perpendicular to their banded structure.

The coals of the highest rank are the *anthracitic*. They are distinguishable from the bituminous coals by their conchoidal fracture and absence of the prismatic cross jointing. Water and volatile matter are extremely low, and nearly all the carbon is fixed carbon.

One of the most important properties of a coal is its heating value, or calorific value, generally measured in British thermal units (Btu*) per pound of coal. Coals are classified by *grade* according to their calorific value, ash content, sulfur content, and ash-softening temperature. A "high-grade coal" is essentially a nearly pure coal.

For technical purposes analyses of coals generally report the following constituents: moisture content, volatile matter (chiefly, but not wholly, combustible matter), fixed carbon, ash, and sulfur.

Lignitic Coals. Lignitic coals are generally chocolate brown, but some are yellowish or black. They range from compact and firm in some varieties (consolidated varieties) to earthy and fragile in others (unconsolidated). Luster is dull and soft to pitchy. Many of these coals show distinctly the texture and grain of wood or intermatting of vegetal fibers; hence comes the name *lignite*. Unconsolidated varieties are called *brown coals*. Hardness ranges from 1 to 2.5; the specific gravity from 0.7 to 1.5. They burn readily with a smoky yellow flame and strong odor. High moisture content is characteristic of lignites. In some it is as high as 40 per cent, that is, the air-dried lignite contains locked up in itself 40 per cent of water.

A lignite from the Glendive district, Montana, has the following composition, which will serve as an illustation:

Moisture	Volatile matter	Fixed carbon	Ash	Sulfur	Heat Value (British thermal units)
34.6	35.3	22.9	7.2	0.7	7090

Lignite occurs in the Cretaceous and early Cenozoic of the Rocky Mountain States, in the Dakotas, and on the Pacific Coast. In Germany, where it occurs in beds up to 300 feet thick, it is a highly valuable fuel resource. Cypress and sequoia are important ingredients in these German coals.

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* A British thermal unit is the heat required to raise 1 pound of water from 62° F to 63° F, 252 calories.

Subbituminous Class. The subbituminous class of coals comprises those in the range between lignitic and bituminous. Subbituminous coal differs from lignite in its black color and its lack of woody texture, and from bituminous coal by its tendency to weather and crumble. Some representative analyses are given in Table 5.

TABLE 5

ANALYSES OF SUBBITUMINOUS COALS

	Moisture	Volatile Matter	Fixed Carbon	Ash	Sulfur	Heat Value (British thermal units)
I....	13.4	32.4	47.6	6.6	0.4	11,120
II....	21.9	34.0	37.4	6.7	0.9	9,070
III....	25.6	27.9	39.2	7.3	0.6	8,290

I. Roundup district, Montana.
II. Glenrock district, Wyoming.
III. Havre district, Montana.

Bituminous class. Bituminous coals are compact, brittle coals of a gray-black to velvet-black color. They have a lamellar, conchoidal, or splintery fracture, and more or less well-defined prismatic jointing. The luster ranges from dull to pitchy; the specific gravity from 1.2 to 1.5. They give a black to brownish-black streak. They burn with a yellow flame and give a strong bituminous odor. Much of the coal shows distinct banding because of the varying luster of the different layers. Generally, no traces of the structures of the constituent plant remains are visible to the unaided eye.

Some bituminous coals fuse or sinter together on heating, leaving a coherent, vesicular residue called coke, and are therefore called coking coals; others fail to do this and fall to powder.

Analyses of bituminous coals are given in Table 6. The sulfur in coal is present partly as organic compounds and partly as pyrite, which is a common impurity. Bituminous coals vary considerably in the relative amounts of fixed carbon to volatile matter, that is, in the proportion of the carbon left behind after the gases are driven off. The gases may amount to as much as 30 to 40 per cent of the original weight of the coal. Coals containing so much volatile matter are called fat coals and are used for making gas and coke. Those with 15 to 20 per cent volatile matter are largely used for steam engines and are often called steam coals. They are transitional to anthracite.

In addition to ordinary coal there are other varieties well defined as to physical characters. *Cannel coal* is dull, lusterless and without structure; it generally shows conchoidal fracture. It is made up largely

TABLE 6

Analyses of Bituminous Coals

	Moisture	Volatile Matter	Fixed Carbon	Ash	Sulfur	Heat Value (British thermal units)
I....	2.8	24.3	66.3	6.6	0.9	14,130
II....	2.4	25.9	66.8	4.9	1.5	14,490
III....	11.9	31.5	49.8	6.8	1.2	11,780

I. Clearfield County, Pennsylvania.
II. Coking coal, Birmingham district, Alabama.
III. Saginaw district, Michigan.

of spores. It is high in gas and burns with a smoky, yellow flame much like a candle flame. The name "cannel" is in fact a dialectal form of "candle." *Jet* is somewhat similar but is characterized by high luster, intense black color, asphaltic appearance, and toughness, which permits of its being readily turned and worked. Its use in mourning jewelry, buttons, and ornaments is well known. It occurs in small, scattered masses in formations, one of the chief sources being Whitby in Yorkshire, England. It is now classified as cannel coal.

Bituminous coal occurs in North America in Nova Scotia; in the Appalachian coal field of western Pennsylvania, Ohio, West Virginia, Kentucky, Tennessee, and Alabama; in the central coal field of Illinois, Indiana, and Kentucky; in Michigan; and in the western field of Iowa, Missouri, Kansas, Arkansas, Oklahoma, and Texas. These coals are of Carboniferous age. In the Rocky Mountain States and on the Pacific Coast there are large deposits of Cretaceous and Cenozoic ages. Elsewhere, in England, Belgium, Germany, France, and Russia and in South Africa, Australia, India, and China, bituminous coal occurs and is mined in great quantities. It is the chief coal of the world. The enormous increase in the annual output in the United States from 137,000,000 tons in 1896 to 620,000,000 tons in 1944, despite competition from oil, gas, and hydroelectric energy, indicates the great industrial importance of bituminous coal.

Anthracitic class. Anthracites are iron black to velvet black. They are brittle and have a strong vitreous to submetallic luster, a more or less pronounced conchoidal fracture, and a hardness of 2 to 2.5. The specific gravity ranges from 1.4 to 1.8. The fixed carbon ranges from 80 to 90 per cent; the volatile hydrocarbons generally do not much exceed 5 per cent; and the remainder of the coal consists of moisture and ash.

Analyses of anthracites are given in Table 7.

TABLE 7

Analyses of Anthracites

	Moisture	Volatile Matter	Fixed Carbon	Ash	Sulfur	Heat Value (British thermal units)
I....	2.8	1.2	88.2	7.8	0.9	13,300
II....	2.2	5.7	86.2	5.9	0.6	13,830
III....	3.0	2.9	86.6	7.5	0.7	13,500
IV....	13.2	2.6	65.3	18.9	0.3	9,310

I. Anthracite region, Pennsylvania.
II. Anthracite region, Pennsylvania.
III. Crested Butte field, Colorado.
IV. Portsmouth district, Rhode Island.

Anthracite requires strong heat to ignite it, and burns with a pale-blue flame, giving great heat without smoke or odor. These qualities and its relative cleanliness particularly adapt it to household use and cause it to command the highest price of any of the coals: it is a "luxury fuel." Actually, its calorific value is less than that of the highest rank of bituminous coals. Some anthracites exhibit on broken surfaces a strong play of spectrum colors produced by iridescent films and are called "peacock" coal.

Anthracite occurs not only in regions of folded strata as previously stated, but also, though not in great quantity, where beds containing bituminous coal or lignite have been invaded by intrusive masses of igneous rock, as in Colorado. The largest deposits of anthracite in the United States are those of eastern Pennsylvania, which contains 99.5 per cent of our domestic reserves of this luxury fuel. Anthracite occurs also in Wales, Belgium, France, Russia, and in the province of Shansi in China, as well as in other places.

Besides the carbonaceous rocks described above, other carbonaceous substances occur, such as graphite, ozokerite or mineral wax, and asphalt and its various modifications, but not in such masses that they can be properly included in a work treating solely of rocks.

CHAPTER XII

ORIGIN AND CLASSIFICATION OF METAMORPHIC ROCKS

Introductory. Metamorphic rocks are rocks that owe their distinctive characters to the transformation of pre-existing rocks. Before their transformation the original rocks may have been sedimentary or igneous, but they have been changed in mineral composition, in texture, or in both, so that their primary characters have been altered or even entirely effaced. Such change or alteration is called metamorphism. Here, as elsewhere in geology, gradations exist—from rocks that are only partly changed to those completely metamorphosed—and no definite line can be drawn between a sedimentary rock on the one hand and its metamorphic product on the other, or between an igneous rock and the metamorphic rock formed from it. Thus limestones grade into marbles, and dolerites merge into hornblende schists, without sharp lines of demarcation. But there comes a point in the transformation of a rock where the characters have become so distinctive either as to mineral composition or as to texture, and generally as to both, that the rock is best regarded as a new kind of rock—a metamorphic rock.

Most rocks are composed of minerals, and most minerals are definite chemical compounds, which as a rule are stable only under definite conditions. If, then, the minerals are subjected as a result of geologic vicissitudes to new conditions, differing from those under which they were formed, and new chemical and physical factors operate on them, they tend to change into other minerals, that is, to become new chemical compounds, which are those that are stable under the new conditions. In short, rocks tend to adjust themselves to their geologic environment. The change in conditions may be so slight that some rock minerals are able to resist it indefinitely, but other less stable minerals succumb. Thus igneous rocks formed by the cooling and solidification of molten magmas may remain deep in the crust for millions of years and on becoming exposed at the Earth's surface as a result of erosion may be entirely unchanged, or with only one or two of their constituent minerals altered. At the Earth's surface they are at once subjected to new conditions: to the combined effects of changes of temperature, moisture, the gases of the atmosphere, and the products of organic life, and they commence to break up and to form new compounds. Their complete conversion is only a matter of time. The same is true of sedimentary rocks, only in lesser degree. They are made up of mineral or rock particles that were deposited in water and were consolidated by pressure and of a cement binding the particles. As long as they stay buried and under fairly stable conditions, they remain unchanged; but when they become exposed to the atmosphere their minerals also tend to decay, especially the minerals that are particularly susceptible to alteration.

The changes that occur at the Earth's surface, however, are not classed as metamorphic changes. All the minerals that are formed by the weathering of rocks on or near the surface, such as the soil minerals, are not described here. They have been previously mentioned under the foregoing rock types from which they are derived, so far as seems to be in the scope of this book. Only those rocks are regarded as metamorphic which, while buried at depth below the surface have undergone changes through the action of certain agents presently to be described and have thus been converted into essentially new kinds of rocks.

Metamorphic agents. The chief metamorphic agents in the Earth's crust are shearing stresses, pressure, the chemical action of liquids and gases, and temperature. In the metamorphism called contact metamorphism (page 148), which is caused by the intrusion of a body of magma, the effect of heat is the most important, that of gases and liquids is less so, and the effects of movements of the crust and pressure are negligible. We shall consider the different agents separately.

METAMORPHISM CAUSED BY SHEARING STRESS UNDER HEAVY CONFINING PRESSURE. Downward pressure due to the weight of the upper part of the Earth's outer crust has little metamorphic effect. It tends to compact the material of sediments by bringing the grains closer together, especially those of the argillaceous sediments, but many sediments, although buried under great thicknesses of deposits for geologic ages, on being raised without disturbance such as folding and on being exposed by erosion have remained practically unchanged except for compaction: they are nonmetamorphic.

It is commonly supposed that the crust is under compression as the result of gradual contraction of the Earth and finds relief from time to time by buckling up into mountain ranges. This compressive force, acting as a lateral thrust, is therefore spoken of as orogenic, *i.e.*, mountain-making. Whole masses of strata and associated igneous rocks—intrusive, extrusive, and fragmental volcanic—are folded and closely appressed by this force. The rocks thus subjected to powerful pressure develop enormous shearing stresses, which tend to produce differential movement between and within their component particles. The deformation by shearing stress is a potent agent in producing metamorphism. Its effects can often be seen megascopically in the large crystals, pebbles, and fossils that have been flattened and elongated ten to fifteen times their original length, or broken into fragments that have been drawn into thin lenses. Shearing stress working alone can produce rocks that show these characteristic effects of metamorphism, without, however, causing in the rocks any change in original mineral composition. In combination with heat and water, it is highly important in bringing about chemical changes and thereby producing new minerals.

Heat. Heat is a powerful metamorphic agent, as is so well shown by the phenomena of contact metamorphism. It greatly increases the thermal agitation of the ions that make up the minerals, and thereby tends to break up the existing chemical compounds, which are the minerals, and to rearrange the ions in the form of new minerals. The heat needed for metamorphism may come from one or more of the following three sources: (1) from the deeper portions of the Earth's outer shell, in which the temperature increases in depth; (2) from the mechanical energy resulting from faulting, folding, and crushing of rock masses; and (3) from bodies of molten magma, which generally invade the crust during and immediately after orogenic revolutions.

Liquids and gases. The chief agent among liquids and gases is, of course, water, which under heat and pressure becomes a powerful chemical agent. It acts as a solvent; it promotes recrystallization. It enters into the chemical composition of some of the minerals, such, for example, as chlorites, micas, amphiboles, and epidote, and is therefore necessary to their formation. It is, without doubt, aided also in its action by the dissolved substances that it may contain, such as alkalies, and by volatile emanations coming from magmatic intrusions, such as boric acid and fluorine, as already explained under contact metamorphism. This activity accounts for the presence in contact-metamorphic rocks of such minerals as tourmaline, chondrodite, and vesuvianite, which are characteristic of certain contact-metamorphic zones, and of micas, hornblende, and other minerals that contain fluorine.

Minerals of metamorphic rocks. Just as certain minerals, of which nepheline and sodalite are examples, are characteristic of igneous rocks, so other minerals, such as kyanite, zoisite, staurolite, and talc, are peculiar to metamorphic rocks. Other minerals, such as quartz, feldspar, amphibole, pyroxene, garnet, and mica, occur in both groups. It should be remembered, however, that the names just mentioned are really names of families, under which several individual mineral species are grouped because they have certain properties in common, such as crystal form. Thus in the amphibole group, arfvedsonite occurs only in igneous rocks; tremolite and uralite occur practically only in metamorphic rocks; and common hornblende occurs in both. Of the pyroxenes, augite is normally at home in igneous rocks and diopside in metamorphic rocks. Of the micas, biotite and muscovite occur in both groups of rocks, though muscovite in igneous rocks is restricted to granites.

In the garnet group, pyrope, the magnesium-aluminum garnet, is formed only in igneous rocks rich in magnesia and low in silica, such as the peridotites; grossularite, the calcium-aluminum garnet, is characteristically at home in metamorphic limestones; and almandite, the ferrous aluminum garnet, occurs in metamorphic argillaceous rocks.

Table 1 gives the minerals that occur in metamorphic rocks. The first column contains those of wide distribution and of prime importance as chief components; the second column, those of lesser importance, which either occur as prominent auxiliary minerals or are locally developed as chief components; and the third, sporadic minerals, which are locally abundant. This classification is true in a broad way, and too much emphasis must not be laid on these divisions.

TABLE 1

MINERALS OF THE METAMORPHIC ROCKS

I	II	III
Quartz	Garnets	Graphite
Feldspars	Staurolite	Tourmaline
Biotite	Epidote	Chondrodite
Sericite and muscovite	Zoisite	Vesuvianite (idocrase)
Amphiboles	Andalusite	Hematite
Calcite	Kyanite	
Dolomite	Sillimanite	
Chlorite	Pyroxenes	
Serpentine	Magnetite	
	Talc	

Chlorite and sericite are especially characteristic of mild (low-rank) metamorphism, whereas kyanite, staurolite, and some of the others are indicative of more intense (high-rank) metamorphism.

Some minerals, notably quartz, can be formed under all conditions of metamorphism or be present as relicts from the original rocks.

Texture of metamorphic rocks. Most metamorphic rocks are obviously crystalline, so that they are frequently referred to as the *crystalline schists*. On the other hand, they resemble sedimentary rocks in having a parallel structure that simulates bedding. This parallel structure expresses itself in a more or less well-defined foliation, by virtue of which the rock tends to split along parallel planes or surfaces determined by the foliation. Phanerocrystalline foliates are called gneisses and schists, according to the degree of perfection of the foliation. Gneisses have a rude or rough foliation; schists have a more closely spaced, well-defined foliation. Although foliation is the outstanding characteristic of metamorphic rocks, a few of them, such as marble and quartzite, are without foliation.

The structure of gneisses and schists is due to arrangement of the minerals into layers of unlike composition or into flat lenses, or to the parallel arrangement of minerals having prismatic or tabular forms, such as hornblende or mica, or to a combination of both causes. It is a result of the shearing to which original rocks were subjected and, under proper conditions, may be imposed on sedimentary and igneous rocks alike. The superficial resemblance that gneisses and schists bear to sedimentary rocks because of their foliation led geologists for a long time to think that all gneisses and schists were derived from sedimentary rocks, and the recognition that many are metamorphosed igneous rocks came only later through petrographic and chemical studies. From the fact that sedimentary rocks of some districts can be traced into metamorphic rocks and these in turn into igneous rocks, it was even thought that the igneous rocks were in part derived from sediments by extreme metamorphism. Such rocks merely represent instances in which both the igneous and the sedimentary rocks have been metamorphosed at the same time, with remnants at either end that are not metamorphosed and whose original characters can therefore be recognized. In the light of our present knowledge we are no more justified in drawing the deduction that igneous rocks are metamorphosed sediments than we are in reversing it and concluding that sedimentary rocks are derived from igneous rocks by metamorphic processes.

VARIETIES OF FOLIATION. Three chief varieties of foliation are recognized:

(1) *Gneissic*, applied to imperfectly or roughly foliated rocks. Gneissic foliation is further characterized as *banded* and *lenticular*. In banded gneisses layers of differing mineral composition are in

parallel arrangement,* as illustrated in Plate 34. This parallel structure resembles bedding, which indeed it may originally have been, but it may also be the result of shear. In lenticular gneisses some of the components are aggregated into thinner or thicker lenses, around which the other minerals tend to be wrapped. Plate 34 shows a lenticular gneiss as seen in cross fracture, *i.e.*, fracture across the foliation. The foliation surfaces of lenticular gneisses are more or less lumpy and tend to obscure and hide the minerals in the lumps. Both layered and lenticular gneisses range from coarse to fine grained. In some layered gneisses the individual layers have lenticular foliation.

(2) *Schistose,* or well foliated. The foliate tends to cleave into thin flakes made up of minerals large enough to be recognized megascopically.

(3) *Slaty.* The mineral grains are extremely small, too small to be seen even with the lens. Slaty foliates are therefore aphanitic. They have the remarkable property of splitting into thin, even slabs, a property particularly well exemplified by roofing slates. Foliation of this kind, because of its perfection, is called slaty cleavage.

There are no hard and fast lines between the three kinds of foliation: slaty cleavage merges into schistosity in the foliates of coarser grain size, and schistosity in turn merges into gneissic foliation.

Metamorphic rocks may contain conspicuously large euhedral crystals of minerals that were formed by the metamorphic processes to which the rocks were subjected.These crystals may be much larger than the average grain size of the rock, and the contrast thus created between them and the matrix, together with their perfect crystal form, produces a strong resemblance to porphyritic igneous rocks. However, such metamorphic rocks are not porphyries, not only because they are not of igneous origin, but also because the large euhedral crystals, the seeming phenocrysts, were not formed first, as are real phenocrysts, but were formed contemporaneously with the minerals of the apparent groundmass in which they lie. The fabric of these rocks is therefore called *pseudoporphyritic* or, more distinctively, *porphyroblastic.* The porphyroblastic crystals, or *porphyroblasts,* frequently contain inclusions of other rock minerals such as bits of quartz and graphite; the inclusions may extend through the large crystals in the lines of original stratification and out beyond them, but this feature can be seen only microscopically. Frequently, even though the pseudophenocrysts are not equidimensional but elongate, they are scattered

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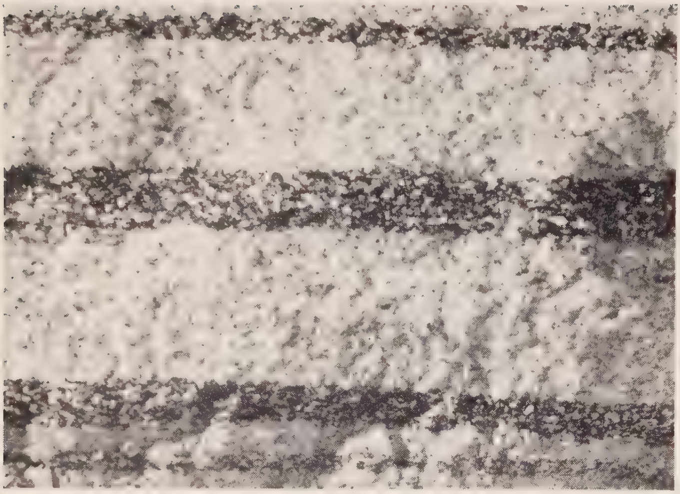
* Layered, a three-dimensional term, is preferable to banded, a two-dimensional term.

PLATE 34.



A. BANDED, OR LAYERED GNEISS, CONNECTICUT.

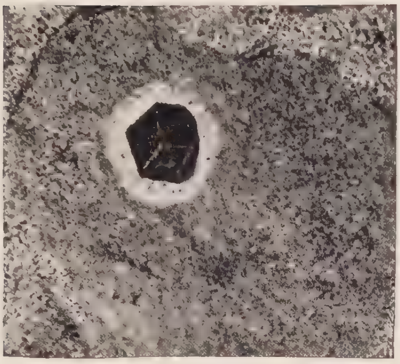
B. LENTICULAR GNEISS.
(Maryland Geological Survey.)

through the rock pointing in all directions; their longer axes do not necessarily lie in the direction of schistosity, as do those of the other minerals, which have become oriented in parallel arrangement during metamorphism. Whether a mineral will grow in the plane of schistosity or in random orientation throughout the rock is probably determined by the relative strength of two tendencies: the inherent directive power of the growing crystal, and the directive effect of the shearing stress simultaneously acting on it. The space in the rock from which material has moved to produce these larger crystals is clearly shown in Plate 35, which contains a photograph of a garnet in gneiss. Around the garnet is a zone of feldspar from which all the ferromagnesian minerals, which are still visible beyond it, have disappeared, having been used up in supplying material to form the garnet.

Porphyroblasts should not be confused with larger crystals and ovoid masses that also give a rock a porphyritic appearance but which are really holdovers from former textures. Such relicts may have been the phenocrysts of a porphyritic igneous rock, or the large grains in some former coarsely granular phanerite, or the pebbles in a conglomerate. All these tend to have ovoid or eyelike forms. They cause the lenticular texture to be conspicuous; it is therefore called *augen texture,* from German *Auge,* an eye. Augen gneisses are widely occurring metamorphic rocks.

RELATION TO PRE-EXISTENT TEXTURES. We find that the characteristic textures just described have been imposed upon rocks in proportion to the degree of metamorphism that they have undergone. But frequently, as though looking through a veil that metamorphism has cast over them, we can see, as it were, beneath these features, the remains of original textures and structures that are characteristic of igneous and sedimentary rocks. Thus, as indicated above, we see that the original texture may have been that of a porphyry, or we may find remnants of the spherulites, lithophysae, and flow lines of felsite lava, or of the amygdules of a basalt. On the other hand, ovoid masses having a different mineral composition from that of their surrounding material may indicate a former conglomerate; parallel layers, differing in general mineral and chemical composition, may indicate former stratified material. Such indications are useful in ascertaining the origin of a metamorphic rock, and sometimes may positively identify it, but deductions should always be made with great caution, for there are many confusing appearances that may lead to error, unless they are checked by microscopic examination and chemical analyses.

PLATE 35.



A. GARNET IN GNEISS, SURROUNDED BY ZONE DEPLETED OF IRON COMPOUNDS.

B. CONTORTED SCHIST.

(U. S. Geological Survey.)

Chemical composition. The chemical composition of metamorphic rocks is extremely varied. Manifestly this must be so, in view of the enormous variety of materials from which they have been formed. If we take them together, as a class of rocks, the composition cannot therefore be as significant as it is in igneous rocks. It may be highly important, however, as an aid in helping to determine their origin. Thus, by examining the chemical analysis of a metamorphic rock, we may find that it is like that of a known igneous rock and can conclude therefore that the metamorphic rock was of igneous origin. On the other hand, the analysis may show that the metamorphic rock could not originally have been an igneous rock and consequently must have been derived from a sedimentary rock.

TABLE 2

Analyses of Metamorphic Rocks

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
I. ..	75.7	13.2	...	1.8	0.4	0.6	2.1	4.7	1.0	...	99.5
II. ..	50.3	14.1	7.0	5.3	7.2	8.1	4.0	2.3	1.6	...	99.9
III. ..	74.7	8.9	9.6	...	1.9	1.1	0.4	1.0	1.1	0.5	99.2

[1] XyO represents small amounts of other oxides.

I. Gneiss, near Freiberg, Saxony.
II. Hornblende schist (amphibolite), Vestana, Sweden.
III. Gneiss, near Rawdon, Quebec.

Among the analyses in Table 2, I might well be that of an ordinary granite, as may be seen by referring to those given under granite; it might also, however, be that of an arkose derived from such a granite. Analysis II has the composition of a gabbro; it might have been such a rock originally, or a dolerite, or basalt; it does not suggest any ordinary sedimentary rock. Analysis III, on the other hand, has no analogy among igneous rocks; the alkalies and alumina are too low for the silica, and the ferric oxide too high; it must be of sedimentary origin and suggests a ferruginous sandstone.

It is inferred, of course, that although the ions have moved through small distances in order to form the new, metamorphic minerals, the chemical composition of the rock mass as a whole has not changed. That this is so, is shown by the fact that innumerable sedimentary rocks, although utterly changed in mineral composition from their former state, still retain their original thicknesses and their continuity. Thus if we focus attention on one bed in a series of strata, perhaps only a fraction of an inch thick, we find that it is sharply marked off by its grain, minerals, and texture from the beds above and below it. There has been no blurring of the bedding planes, there has been no melting and no transfusion of substance. Consequently the changes that occurred were, so to speak, internal within the individual beds. To the general statement that the bulk composition of the rock does not change during metamorphism, there is one exception, namely that volatile substances may be driven out and, conversely, new ones may come in and enter into new mineral combinations, as previously explained in the description of the action of liquids and gases as metamorphic agents. This change is most strikingly exemplified in the metamorphism of impure limestones, as described in the section dealing with marble (page 327) and is thoroughly analogous to what has already been stated under contact metamorphism.

Injection gneisses and schists. As was previously mentioned, part of the heat of metamorphism and part of the liquids and gases involved in forming new minerals, are supplied by intrusive magmas, which are particularly likely to invade those portions of the crust where mountain-making forces are causing the metamorphic agents to become active. In such areas the effect of contact metamorphism becomes so widespread that no definite line can be drawn between contact and regional metamorphism. Indeed, the earlier-formed intrusions may themselves become more or less metamorphosed, or have metamorphic textures imposed upon them by repetitions of the processes, while they are in a solid, or yet partly plastic, condition.

These magmas, rising under great pressure into rocks already schistose may produce another effect: they may insinuate themselves as thin dikes, sills, and lenticles into the schists surrounding them, so that composite rocks are formed, partly igneous and partly metamorphic in composition. As previously explained under contact metamorphism and pegmatite dikes, these effects may be greatly intensified by liquid and gaseous emanations from the magma masses. This process is termed *injection* of schists by magmatic material, and the products are *injection gneisses*.

Occurrence and age. Metamorphic rocks are widely distributed over the Earth. Moreover, in many places they occupy great areas throughout which they are the only rocks exposed. Their extension over such large regions has given to the processes that produced them the name of *regional metamorphism*, to signalize the contrast between these zones and the relatively small metamorphic zones around intrusive igneous masses.

Metamorphic rocks occur also in folded mountain ranges, of which they form the core, which subsequent deep erosion has brought to light. This fact is so well established that when we find areas where the rocks are intricately folded and completely metamorphic but do not form a mountain mass, we conclude that a mountain formerly existed but has been eroded away. In short, metamorphic rocks become exposed to view at the surface only by denudation.

These facts led to the view, formerly held, that metamorphic rocks must be of very great geologic age. This is by no means true, however. For on the one hand we find unmodified sands of Cambrian age in eastern Russia, and unaltered beds of Ordovician age in the upper Mississippi valley, which have remained geologically undisturbed; on the other hand strongly folded strata of Mesozoic and Cenozoic age in the Alps and other mountain ranges are in places profoundly metamorphosed.

Classification of metamorphic rocks. It would be logical to classify metamorphic rocks according to the origin of their material, and to separate those of igneous from those of sedimentary derivation. For some this distinction can be made; thus marble is obviously not of igneous origin. However, when we attempt to carry this principle through, it quickly becomes impracticable, especially if we are limited to megascopic means of determination.

TABLE 3

Analyses of Arkose and Granite Porphyry

SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O, etc.	Total
69.9	13.1	2.5		trace	3.1	5.4	3.3	1.0	98.3
69.9	14.9	1.8	0.6	0.6	1.5	5.3	3.9	1.3	99.8

For instance, in Table 3 the upper analysis is that of the Triassic sandstone of Portland, Connecticut, a fine-grained arkose full of feldspar; the lower one is that of a granite porphyry from the Crazy Mountains in Montana. If these two rocks, one sedimentary and the other igeous, had become so thoroughly metamorphosed that they had lost all traces of their original textures, they could not be distinguished apart and what they were originally could not be determined. Tables 4 and 5 show that gneisses and schists are derived from various kinds of rocks.

On the basis of the classification of the sedimentary rocks previously given, the relation between the most common sedimentary rocks and their metamorphic derivatives is shown in a general way in Table 4.

TABLE 4

SEDIMENTS AND DERIVED SEDIMENTARY AND METAMORPHIC ROCKS

Sediments	*Sedimentary Rocks*	*Metamorphic Rocks*
Gravel	Conglomerate	Gneiss, and various schists
Sand	Sandstone	Quartzite, and various schists
Silt and Clay	Shale	Slate, phyllite, and various schists
Limy deposits	Limestone	Marble

The igneous rocks can be roughly divided into three groups: (1) composed chiefly of feldspathic minerals; (2) composed mainly of ferromagnesian minerals; and (3) composed wholly of ferromagnesian minerals. The relation between them and their metamorphic derivatives is shown in a general way in Table 5.

TABLE 5

IGNEOUS ROCKS AND THEIR METAMORPHIC DERIVATIVES

Igneous Rocks	*Metamorphic Rocks*
Coarse-grained feldspathic types, such as granite	Gneiss, schists, and phyllites
Fine-grained feldspathic types, such as felsite and tuff.	Schists and phyllites
Ferromagnesian rocks, such as dolerite and basalt.	Hornblende schists
Ultramafic rocks, such as peridotite and pyroxenite.	Serpentine and talc schist

Another method of classification, developed chiefly by Swiss and Austrian geologists, disregards the origin of the material entirely and considers only the chemical composition of the rocks. In this classification, the metamorphic rocks are divided into twelve groups of different chemical composition. The Earth's crust is divided into three zones—an upper, middle, and lower—in which temperature and pressure increase progressively downward. The effect on each group of the metamorphism in these depth zones is considered. Material of a given composition is found to have yielded rocks that differ in mineral composition and texture according to the zone in which the metamorphism occurred. The major classes are based on chemical composition, whereas the subdivisions are based on mineral composition and metamorphic condition, and in this way the manifold rocks are classified.

From this system has evolved the idea of classifying metamorphic rocks on the basis of intensity of metamorphism. Rocks of three ranks (or grades) of metamorphism are recognized: (1) low rank; (2) middle rank; (3) high rank. Certain index minerals are used to determine the rank of the rocks in which they occur. For example, rocks containing sericite and chlorite are low-rank metamorphic rocks.

The practical purposes of field work and megascopic determination oblige us at present to classify the metamorphic rocks arbitrarily according to their evident mineral composition or texture, or a combination of both. Stress is laid on either mineral composition, or, as in the slates, on the texture, whichever is the more evident and characteristic.

In agreement with this, metamorphic rocks are classified as in Table 6. A more detailed table is given on page 334.

TABLE 6

MAIN GROUPS OF METAMORPHIC ROCKS

I. NONFOLIATES
    Hornfelses
    Tactites
    Mylonites
    Quartzites (metaquartzites)
    Marbles

II. FOLIATES
    Gneisses
    Schists
    Phyllites
    Slates

CHAPTER XIII

DESCRIPTION OF METAMORPHIC ROCKS

MYLONITE

Mylonites are aphanitic rocks of flintlike, felsitic, or, if streaky, rhyolitic appearance. Many of them have an obscure foliation, generally best seen on weather-etched surfaces. They occur along the base of some (but by no means all) overthrust sheets and as dikelike masses that have been injected into the overlying rocks. The mylonite zones along the base of overthrust sheets range from a few feet to many hundred feet in thickness.

In the formation of a mylonite, the constituent minerals of the original rock were mashed, pulverized, and dragged out; the result is a streaky or banded, compact flinty rock, called a *mylonite* (Greek *mylon*, a mill), because the rock from which it was formed has been through the metamorphic mill. The words "mashing" and "pulverizing" emphatically call attention to the most striking feature of mylonitization, namely the production of an immense number of minute grains at the expense of the original grains in the rocks mylonitized, but they are somewhat misleading, inasmuch as mylonites are compact, strong rocks.

Rocks as strong as granites have been reduced to mylonites. In the field favorable exposures may show the transition from undamaged granite to mylonite. If such a transition is not visible, the origin of the mylonite may remain enigmatic, so completely does the mylonite differ from the rock from which it was derived. In the intermediate stages of reduction from granite to mylonite the mylonitic base contains angular fragments of feldspar and quartz and the whole resembles arkose, but as these angular fragments become less in number, the mylonite may resemble a rhyolite containing phenocrysts of feldspar and quartz. Rock in which all the particles have been reduced and made imperceptible to the eye is called *ultramylonite*. In an ultramylonite, the constituent particles are so minute that X-ray analysis is required to show that they are crystalline mineral grains.

The nature of a mylonite is, of course, influenced by the mineral composition of the rock from which it is derived. If, for example, a granite containing considerable biotite is mylonitized, the biotite is smeared out into an immense number of flakelets, all in parallel arrangement, and the resulting rock resembles a black slate or a phyllite.

Mylonites are the products of intense mechanical metamorphism: they are the finest examples of metamorphism effected by purely mechanical processes, uncomplicated by the growth of new minerals. However, transitional rocks occur that make a complete sequence from mylonite of purely mechanical derivation to rock in which some new minerals have formed, and finally to rock in which all the minerals are of new growth and consequently all evidence of the milling through which the rock went has disappeared. Such a transitional sequence supports the current theory that many metamorphic foliates have been formed by the growth of new minerals during forced flowage of rocks in the solid state.

GNEISS

Gneiss denotes a rock characterized by a roughly developed, imperfect foliation. Thus when we say "granite gneiss," "syenite gneiss," or "diorite gneiss," we are referring to a rock that has the composition indicated by the first word and the texture indicated by the second. The term gneiss includes extraordinarily diverse rocks—from feebly metamorphosed rocks to the most highly metamorphosed rocks. Gneiss is a descriptive, not a genetic term. Gneisses derived from sedimentary rocks are called *paragneisses*; those derived from igneous rocks are called *orthogneisses*.

Mineral composition. Feldspars of various kinds, including potassium feldspars and plagioclase, occur in gneisses, but they can rarely be distinguished by megascopic means. The feldspars are white to gray, or reddish, as in granite, and generally occur as more or less round or elongate, lenticular grains. This lack of definite form makes them more difficult to distinguish from the quartz than in most granites, and consequently the mineral cleavage that is characteristic of feldspar should be carefully sought. Large feldspar grains, the size of a pea or even larger, occur in some gneisses, giving them a porphyritic appearance; if these grains are examined against the light for their cleavage surfaces, many are seen to be Carlsbad twins. These large crystals may formerly have been phenocrysts in the porphyritic granite from which the gneiss was derived, they may have been feldspar pebbles in a conglomerate or arkose, or they may have been made by injected material.

The quartz also occurs in more or less round grains or lenticles, or in granular aggregates with the feldspar. It is white or gray; less commonly it is bluish. In the larger grains it is recognizable by its glassy luster and conchoidal fracture.

The mica is biotite, muscovite, or both. Biotite is black or dark brown; muscovite is white, yellowish to light brown, or pale green. The micas in gneisses have no distinct crystal form but are in flakes or irregular leaves, drawn out in bands, or in thin patches. They are usually distributed along the structure planes of the rock; and the easy cleavage of the micas, thus arranged in one direction, largely determines the foliation of the gneiss and emphasizes the gneissic texture. Consequently the surface of easy fracture of a gneiss may appear to be coated chiefly with mica, and, judging from this appearance alone, one is likely to form an exaggerated idea of how much mica occurs in the gneiss; therefore the surface of cross fracture should always be examined also, in order to determine how much mica is really present. This seeming predominance of mica in foliates is especially marked in mica schists and in those gneisses whose large content of mica makes them transitions into mica schists.

Hornblende occurs in gneisses; in some it is associated with biotite, but in others it occurs alone, forming a special variety. It occurs as dark, prismatic crystals without good terminations. Minute crystals may be aggregated into flattish lumps and layers.

Many other minerals occur in gneisses; if abundant they determine special varieties of gneisses. Garnet, of the dark-red variety, almandite, occurs perhaps the most generally. The crystals of garnet may be conspicuously large compared with the other constituents. Epidote also may occur, as may *graphite,* in some varieties. Sillimanite, a mineral of the same composition as andalusite and kyanite, occurs in some gneisses in bundles and brushlike groups of slender fibers. Tourmaline occurs in notable quantity only where the gneisses were subjected to pneumatolysis. In some gneisses chlorite is the dark constituent instead of mica.

Texture. The essence of the gneissic texture is a rough parallel arrangement of the constituent minerals, best indicated by platy and columnar minerals, commonly mica or hornblende. If the amount of mica or hornblende is small, the gneissic foliation is but poorly developed; the more the mica, the more distinct the gneissic foliation. Some layers are thick and coarse, giving a pronounced banded effect; some are extremely thin. Gneisses may have a layered structure as well as a foliation. The individual layers in some gneisses persist for considerable distances; in others they are short and lenticular. The gneissic foliation of some gneisses is so faint as to be scarcely perceptible in the hand specimen but is clearly evident on large exposed surfaces of the rock.

Gneissic banding may extend for long distances as even regular layers, and such persistent regular layering is generally an inheritance from sedimentary bedding. The layering may be curved, folded, and contorted in the most complex and remarkable manner, so that the folds range from large convolutions down to minute or even microscopic crinkles. Intricately folded gneisses are shown on Plates 35 and 36.

Some gneisses, like some granites, are porphyritic, as the result of containing large, well-defined crystals of feldspar. These gneisses as a rule were originally porphyritic granites and have had a gneissic foliation impressed on them. The large crystals in some gneisses, however, were formed by growth from introduced material, *i.e.*, by metasomatic replacement. Such gneisses are allied to and may grade into varieties that have a short, thick, lenticular structure and contain ovoid masses of feldspar or quartz. The ovoid bodies are called augen (German, *Augen,* eyes) and the rocks containing them are called *augen gneisses.* As was explained previously, the augen may have formed in several different ways.

In some gneisses pebbles of various kinds can be seen. These pebbles were derived from pre-existing masses of granite, quartzite, and other kinds of rocks. They are generally in the shape of flat lenses, but that they were originally pebbles is evident, and the gneiss was manifestly once a conglomerate, the finer material of which has reorganized into new, metamorphic minerals, whereas the larger pebbles have remained mostly unchanged, save in shape.

"Primary gneisses" are foliated granites whose foliation was produced by flowage while they were consolidating from the magma. During this flowage the already formed minerals suspended in the still liquid portion of the magma took on a parallel arrangement, especially the biotite. Such gneisses are therefore not metamorphic rocks at all; hence they are called primary gneisses.

Color. The color of gneisses is too varied to be a diagnostic feature. It depends on the color of the quartz, feldspar, and other light-colored minerals, and on the proportion of all these to the amount of biotite or other dark-colored minerals in the gneiss. Also, in paragneisses (gneisses of sedimentary origin) carbonaceous material may be present as graphite, which imparts a dark color. Hence, gneisses range from almost white through light gray or red into darker shades and into brown, green, and even black.

PLATE 36.

CONTORTED GNEISS, OTTAWA RIVER, CANADA.
(After R. W. Ells.)

Chemical composition. Because gneisses have been derived from rocks of many different kinds, they have such a highly variable chemical composition that composition is not a distinctive feature. The many gneisses that consist largely of quartz and feldspar contain silica, alumina, and alkalies, and generally more or less iron and lime, but these oxides range within wide limits. Table 1 gives analyses of a few typical gneisses. In spite of the uncertainty involved, the analysis of a gneiss may suggest a clue as to its origin. Thus analyses III, IV, and VI in Table 1 are quite unlike those of any igneous rock and are almost certainly analyses of material of sedimentary origin, whereas the other analyses may be of rocks of igneous derivation.

TABLE 1

ANALYSES OF GNEISSES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
I...	71.0	15.0	1.1	1.8	0.7	0.3	2.5	5.8	1.1	0.5	99.8
II...	65.1	16.4	0.9	5.6	2.4	2.4	3.3	1.9	0.7	1.3	100.0
III...	78.3	10.0	1.8	1.8	1.0	1.7	2.7	1.3	1.0	0.9	100.5
IV...	82.4	11.3	1.0	0.3	0.2	0.2	0.6	1.0	2.5	0.2	99.7
V...	52.2	18.8	2.7	5.3	5.1	8.0	3.3	1.6	1.4	1.7	100.1
VI...	44.5	17.5	3.4	12.6	5.7	3.3	3.6	3.5	1.6	5.2	100.9

[1] XyO represents small amounts of other oxides.

I. Granite gneiss, Lincoln, Vermont.
II. Garnet-biotite gneiss, Fort Ann, Washington Co., New York.
III. Gneiss, fine grained, Great Falls of Potomac River, Maryland.
IV. Schistose gneiss, Marquette region, Michigan.
V. Plagioclase gneiss, Mokelumne River, California.
VI. Gneiss (kinzigite), Schenkenzell, Black Forest, Baden.

Varieties of gneiss. Many varieties of gneiss are distinguished by geologists and petrographers. They are based partly on differences in texture, which distinguish "banded gneiss," "lenticular gneiss," and "augen gneiss"; partly on the presence of characteristic or unusual minerals which distinguish "biotite gneiss," "hornblende gneiss," and "epidote gneiss"; and partly on general composition, which distinguishes "granite gneiss" and "diorite gneiss." In the last usage, as already explained, gneiss is used in the sense of a textural modifier.

In this book it is not feasible to describe all the kinds of gneisses, but a few of the most prominent are mentioned. Among mineralogic varieties, by *common gneiss*, or *gneiss* for short, *mica gneiss* is meant. If further distinction is required, the kind of mica can be specified, and accordingly we have *biotite gneiss, muscovite gneiss*, and *biotite-*

muscovite gneiss. If the mica is accompanied or proxied by some prominent mineral, as is common, other varieties of gneisses can be distinguished, such as *hornblende gneiss*, *epidote gneiss*, and *garnet-biotite gneiss*. Among the varieties of gneiss that are based on general composition, all the feldspathic phanerites occur as gneisses having well-marked foliation. Accordingly, we have *granite gneiss* (by far the most common variety), *syenite gneiss*, *diorite gneiss*, and even *gabbro* and *anorthosite gneisses*. The gneissic foliation was imposed on some of these rocks after they had solidified, as the result of flowage in the solid state under a heavy confining pressure, and on other gneisses by magmatic flowage while they were still mushy, or only partly solidified.

Although the study of thin sections under the microscope is often of great assistance in the endeavor to ascertain the origin of a gneiss and thus to understand its relation to other rocks better, it is by no means always necessary. Much can be done in the field by obtaining all the facts that can be ascertained and by carefully correlating these facts with one another. From place to place the gneisses should be studied with the lens, and all changes in mineral composition or texture should be noted. Some of the chief points to be looked for in distinguishing gneisses of igneous origin from those of sedimentary origin are the following. Igneous gneisses are likely to have a more uniform composition and texture over large areas. The zone bordering the contact with other rocks may show the remains of a former endogene contact zone, such as a fine-grained border, a porphyritic border, or pneumatolytic minerals, among which tourmaline is especially likely to occur. The remains of former aplite dikes or pegmatite veins, as described above, also are testimony of a former contact zone. The enveloping or bordering rocks may show remains of a former aureole of contact metamorphism by a change in mineral composition or texture, and by the presence of tourmaline or other pneumatolytic minerals. The plane of contact of the gneiss and its neighboring rocks should be examined to see whether they are interwoven, as contorted interlaminated beds might be expected to be, or whether the gneiss cuts across them. In fact, all the field characters indicative of intrusion, which have been described under granite, should be looked for under the veil that metamorphism has cast over the region. They may, of course, have been entirely obliterated, but some may have persisted and be valuable clues.

In sedimentary gneisses, on the other hand, more abrupt changes in composition and texture can be expected from place to place, both on a large and on a minute scale. The remains of former pebbles, or small, lenticular masses of different composition indicative of pebbles, should be sought for. The presence of carbonaceous matter, or graphite, diffused through the rock, or collected in spots or streaks, also indicates a sedimentary origin. The absence of any of the signs of intrusion, and the character of the contact, as mentioned above, may likewise be clues.

If, in addition to the facts observed in the field, a chemical analysis of a well-selected specimen, or series of specimens, of the gneiss can be obtained, this additional information may prove to be of great value.

General properties and uses of gneiss. Many gneisses are solid and massive rocks and their general properties closely resemble those of massive igneous rocks. Thus granite gneiss closely resembles granite and is used in the same manner for building and structural purposes. But many granite gneisses contain so much mica that they have too marked a foliation to be of much value. In general a gneiss should be so placed that its plane of easiest fracture lies in the mortar bed with the cross fracture exposed; otherwise it is likely, like some sedimentary rocks, to split and scale badly. The jointing and erosion forms of granite gneiss are, in general, like those described under granite. The weathering is also similar, and gneisses form fertile sandy soils which pass into loamy ones, as the decay of the feldspar and its alteration into kaolinite become more complete.

Occurrence of gneiss. Gneiss, especially mica gneiss, is one of the most common and widely distributed of metamorphic rocks. The Precambrian rocks of many regions are almost entirely composed of gneiss, and to attempt to mention all the localities at which gneiss occurs would be practically equivalent to a description of the occurrence of the world's Precambrian. Gneisses are not restricted to the Precambrian, however; they occur in later formations, into the Mesozoic. Gneisses occur throughout New England, and southward along the Piedmont Province into Georgia. They occur in the Adirondacks, in the Rocky Mountain region, and in other places in the United States. They cover large parts of eastern Canada and are prominent in Scotland, Norway and Sweden, Finland, parts of Germany, and the Alps.

Granulite. Associated with gneisses at some localities is a moderately foliated rock composed almost wholly of quartz and feldspar. It is nearly, or wholly, free from mica, and is usually of fine to minute grain size, so that, except for its schistose character and associated rocks, it is much like an igneous felsite or aplite. It commonly carries minute red garnets, as well as small quantities of other minerals, notably kyanite. Chemically, it is similar in composition to some felsites or the aplite variety of granite, and it probably represents in general igneous rock of this nature that has been involved in metamorphic processes. Granulites of this kind occur in Saxony and other places in Germany, where they were first studied.

The foregoing account of granulite is based on German usage. In British and American usage, however, granulite is commonly used as a name for fine-grained phanerocrystalline metamorphic rocks in which foliation is weak or absent. Diopside granulites, for example, are described as occurring in New Hampshire.

MICA SCHIST

Mica schist is closely related on the one hand to gneiss and on the other to quartzite. It is not only a common companion of gneisses in regions of metamorphic rocks, but in many places gneiss grades into mica schist, so that no definite line can be drawn between them. Of the great class of rocks known as schists, mica schist is the most widely distributed and important.

Composition. Minerals and texture. The essential minerals of mica schist are quartz and mica, and the mica especially gives the rock its particular character. The most common micas are silvery white muscovite and biotite of a dark color. The sodium-bearing mica (paragonite) is extremely rare. Muscovite and biotite occur with the quartz either singly or together. The micas are in irregular leaves or tablets, without crystal boundaries, or in leafy or foliated aggregates. Biotite and muscovite are intergrown, generally with a common cleavage. The micas lie with their cleavage planes in the plane or surface of schistosity, and this parallel arrangement of the mica plates produces the extraordinarily fissile character of the rock. They are also very often curved, bent, or twisted, as may be easily seen by the reflections from their cleavage surface. The cleavage of the mica is so marked that the foliation surface, the plane of schistosity of the rock, appears to be completely coated by mica. The impression may therefore be that mica is the only mineral present; in order to see the quartz, the other essential component, the cross fracture should be examined with the lens. The quartz forms irregular grains, or aggregates of grains, and these are arranged in small lenses, or in thin layers, parallel with the layers of mica.

Mica schists, though generally composed of quartz and mica alone, may carry prominent crystals of other minerals, some of them of large size. The most common of these is the dark-red garnet, almandite, sparsely or thickly scattered through the schist, and ranging in size from that of coarse shot to that of a plum. These garnets may be in the form of roundish nodules, but as a rule they have more or less distinct crystal form and are beautifully crystallized as dodecahedrons and trapezohedrons. Garnetiferous mica schist of this kind is a widely distributed, striking metamorphic rock.

Other minerals that occur in mica schist in a manner similar to garnet are staurolite, kyanite, sillimanite, and hornblende. These minerals may form large well-shaped crystals, especially the staurolite and kyanite. Graphite occurs in some mica schists in quantity sufficient to produce a distinct variety. Graphite is such a strong coloring matter that a relatively small amount causes the schist to seem to be almost entirely composed of it. In consequence, unsuccessful attempts have been made in places to exploit such lean schists for graphite.

Hornblende, if present, is in dark-colored prisms. By increase in its amount transitions are made into hornblende schist.

Kyanite, sillimanite, and staurolite may occur in prismatic crystals as much as several inches in length. They were formed contemporaneously with the metamorphism of the rock, and they produce the porphyroblastic texture explained on page 202. Another variety of mica schist contains more or less calcite mingled with the quartz; it is readily detected by its effervescence with acids. This variety commonly contains accessory garnet, epidote, and hornblende, *i.e.*, high-calcium minerals.

The well-developed foliation of mica schist is its special feature, and its ready fissility is produced by the mica. If the components are in thin, parallel layers, the foliation surface is smooth and flat; if lenticular arrangement of the quartz is prominent, the surface is uneven or lumpy. Frequently the surfaces of schistosity are folded and crumpled, showing that the schist was again subjected to deformation after its metamorphism.

Chemical composition. The chemical composition of mica schists, like that of gneisses, is too variable to be of specific value, because of the diversity of composition of the sediments from which they were formed. Clearly, however, they must contain silica, alumina, and potash to form the quartz and mica, and also magnesia and iron, if biotite is present. The fact that there is more magnesia than lime, along with high content of silica, is a character foreign to igneous rocks and clearly indicates sedimentary origin. Most mica schists were formed by the metamorphism of argillaceous rocks. Two analyses of typical mica schists are given in Table 2.

TABLE 2

Analyses of Mica Schists

SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
64.7	16.4	1.8	3.8	3.0	0.1	0.1	5.6	3.1	0.8	99.4
64.8	14.4	1.8	4.5	2.3	2.3	1.4	5.0	2.0	1.9	100.4

[1] XyO represents small amounts of other oxides.

General properties. The color of mica schists ranges from silvery white through gray, yellow, or brown tones into very dark, depending on the proportions of light and dark mica and the amount of carbonaceous material. Pure muscovite schists are silvery white or light gray. The hardness and firmness of mica schists depend on the proportion of mica; the more mica is present, the softer and the easier to cleave they are. For this reason highly micaceous schists have little or no value as building stones. Some of the firmer mica schists, however, make striking building stones; the buildings of Princeton University are notable examples. Veinlets and lenticular masses of quartz that fill cracks and cavities opened by movement and foldings of the rocks are common.

During the weathering of mica schist, the varieties composed of muscovite are in the main mechanically distintegrated without much chemical change. Muscovite strongly resists alteration by weathering, and consequently soils formed from muscovitic mica schists are filled with its sparkling flakes. The biotite in mica schists weathers easily; the schists turn yellow or brown and lose their luster, and eventually much limonite forms and permeates the rocks.

Varieties and occurrence. Muscovite-biotite schist containing both micas, as its name implies, with associated garnet, is the most usual kind and occurs all over the world abundantly in metamorphic regions. It is generally associated with gneisses. It covers large areas in New England and extends southward to Georgia. Staurolitic mica schist occurs in many places in New England and in the Piedmont Province. Kyanite-mica schist occurs in many places in the Eastern Atlantic States. A variety in which the mica is paragonite comes from the Saint Gotthard region in the Alps and is seen in mineral collections on account of the beautiful crystals of kyanite it contains. Hornblendic mica schists occur as included lenticular masses, some of large dimensions, in various places, in the ordinary mica schists.

Graphitic mica schist occurs in Madagascar, where it is an important commercial source of graphite; in Pennsylvania and Alabama where it was a war-time source of graphite; and in many other metamorphic regions.

An interesting variety is conglomerate mica schist; it contains pebbles of quartz, granite, and other rocks, which commonly are flattened, lenticular, or drawn out by the shearing to which they were subjected. It is closely related to the conglomerate gneiss previously described and is of similar origin.

Transitions and relation of other rocks. Paragneisses, the gneisses formed from sediments, and the mica schists have been derived from similar rocks: from feldspathic sandstones, shales, and conglomerates. During the transformation to mica schists the feldspar was converted into mica; in the gneisses it has mostly persisted or has recrystallized. This statement is not intended to affirm that this is the only origin for mica schists, but that it is the most usual one; some have been formed from quartzose feldspathic igneous rocks. On the whole, gneisses have most probably been formed from the conglomerates and coarser-grained sandstones, and the mica schists have been formed from the finer-grained sandstones and from shales.

As previously stated, no hard and fast line can be drawn between gneisses and mica schists. Whether a given foliate in the transitional range should be classed as gneiss or mica schist is often a matter of opinion. In general, if the foliate contains sufficient mica to be conspicuously schistose, it is best classified as a mica schist; but if it contains a small amount of mica and is imperfectly foliated, it is best classified as a gneiss.

QUARTZITE

Quartzite is a firm, compact rock, composed of grains of quartz so firmly united that fracture takes place across the grains instead of around them. It is in general a metamorphosed sandstone, and although no hard and fast line can be drawn between sandstones and quartzites, since all degrees of transition exist between them, quartzites are far more coherent than sandstones. Sandstones have a more or less sugar-granular feeling and appearance, and the individual grains are distinctly perceptible to the eye or lens. In quartzites the fractured surface is uneven, splintery, or conchoidal; the luster is vitreous like that of quartz, and the grains are imperceptible or nearly so. The difference arises chiefly from the fact that when sandstone is broken the fracture takes place in the cement, leaving the grains unbroken and outstanding, whereas in quartzite, as just mentioned, the fracture takes place through the cement and the grains alike. This difference serves as a practical distinction between the two rocks.

Minerals and general properties. Pure quartzites consist entirely of quartz, but most quartzites carry other minerals in greater or less abundance. These auxiliary minerals may be detrital grains, such as feldspar, mixed with grains of quartz, or new minerals formed by the metamorphism of the clay or calcareous cement that formerly filled the interstices between the grains of the sandstone. Such newly formed minerals are muscovite, biotite, kyanite, and epidote. Iron hydroxides are converted by metamorphism into magnetite or hematite, and carbonaceous substance is altered into graphite. From the kind of minerals formed during metamorphism, the metamorphic rank of the quartzite can be determined.

The metamorphic minerals may be of microscopic size and may give the rock a distinct color—green, blue, purple, or black; but in many quartzites they are large enough to be clearly seen with the lens. The most common are muscovite and biotite, which, as they increase in amount, give the rock a more and more schistose character, through which it attains a capacity for cleavage along the planes of the mica. Abundance of mica produces rocks that are transitional into mica schist. Quartzite, mica schist, and gneiss thus form a continuous series in which the divisional lines are purely arbitrary.

Quartzite is generally white, light gray or yellowish to brown, but these colors are modified by included material acting as a pigment, as explained above. The jointing of quartzite is usually platy but may be widely spaced. Widely jointed quartzites are quarried in some places and furnish good material for structural purposes.

The chemical compostion of pure quartzite is nearly that of silica alone, but if more or less clay or calcareous material was mixed with the sand, small amounts of alumina, iron, lime, and alkalies appear. This contrast is illustrated in the two analyses quoted in Table 3.

TABLE 3

Analyses of Quartzites

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	Total
I....	97.71	1.39	1.25	....	0.13	0.18	....	....	....	100.66
II....	74.22	10.61	7.45	0.85	1.48	0.56	2.12	1.08	1.79	100.16

I. Pure quartzite, Chickies Station, Pennsylvania.

II. Impure quartzite, Pigeon Point, Minnesota, contains small quantities of feldspar, mica, chlorite, and magnetite.

Varieties. Varieties of quartzite are chiefly those differentiated by the presence of some included substance. Thus we have epidotic quartzite, graphitic quartzite, sillimanite quartzite, and many others. Quartzite may have been mechanically deformed by shearing, which has crushed the original grains and impressed on the quartzite a more or less well-defined schistose structure. Some quartzites contain pebbles, of different sizes, which may have retained their original shape or may have been deformed by solid flowage to ovoid, lenticular, or long cylindrical bodies. These remarkable rocks are called *conglomerate-quartzites*, and like conglomerate gneisses and conglomerate-mica schists they were formed from what were originally gravels.

Buhrstone is a variety of quartzite that is full of long hollows or pores. Notwithstanding its porosity, it is quite firm; its hardness and toughness have caused it to be made into millstones. It is thought to have been originally a more or less pure limestone filled with fossils, which has been replaced by the action of solutions containing silica and has been converted into a quartzite consisting mostly of chalcedony; the cavities represent the leached-out fossils.

Occurences. Quartzite is widely distributed, mostly among metamorphosed strata. It is common in eastern North America, in the Rocky Mountain Cordillera, and in many other parts of the world. The occurrence of some special varieties has already been mentioned.

Alteration. On account of the insoluble, unyielding nature of its constituent grains and their cement, quartzite strongly resists erosion and atmospheric agents. Where it is abundant in mountain regions and areas that are being eroded, it forms prominent features of the landscape, bold ledges, cliffs, castellated crags, and spires. Eventually, quartzite breaks down into sandy soil of poor quality.

Distinction from other rocks. Quartzites, which appear to be very homogeneous, may be confused in the outcrop or hand specimen with some limestones or felsites of similar color and texture. They are easily distinguished from limestones because they are far harder and do not effervesce with acid; they can be distinguished from felsites by blowpipe tests and possibly by their different mode of geologic occurrence. It will be recalled that the chief minerals composing quartzite, limestone, and felsite are quartz, calcite, and feldspar, respectively; the rocks should be tested accordingly.

SLATE

Slates are homogeneous rocks, of such fine grain that the individual constituent mineral particles cannot be distinguished by the unaided eye or lens, and characterized by a remarkable planar cleavage, by means of which they can be split readily into broad, thin sheets.

Slates are low-rank metamorphic rocks: they represent, in mildly metamorphosed condition, the finest material of the land waste produced by erosion, the same material that, among the unmetamorphosed stratified rocks, appears as clay and shales of various kinds. With such material more or less volcanic dust and ash may have become mingled.

Mineral composition and other properties. Because the mineral particles in slates are so minute, the composition cannot be determined megascopically. However, since slates have been formed from clays derived from many sources, microscopic and X-ray analyses detect in them many minerals, chiefly quartz (30 to 45 per cent), sericite (40 per cent), chlorite (5 to 20 per cent), and carbonaceous matter. Sericite is the most distinctive component and occurs as minute laths, about 6 to 30 microns long (a micron = 0.001 millimeter) and 1 micron thick. Kaolinite, which one might naturally expect, appears to be restricted to the very lowest-rank slates, in which metamorphism was arrested at a very early stage. Slates may contain crystals of pyrite, which can readily be seen with the eye or lens and which may attain large size, some as distinct crystals, and some as concretions or as replacements of fossils. Marcasite also has been found in slates. Veins and lenses of secondary quartz are common in them, those of calcite more rare.

The color is chiefly gray, dark gray, or black, according to the amount of graphite or carbonaceous matter; it is green if chlorite is abundant, and red, purple, yellow, or brown from oxides of iron. The surface of the slaty cleavage is more or less lustrous; but the surface of cross fracture is dull, without luster. Although slate is firm and not friable, it is rather soft, so that it can be readily cut, a feature of great technologic value. The specific gravity of the average slate is about 2.8.

The chemical composition is shown in Table 4, in analyses of typical slates made in the laboratory of the U. S. Geological Survey. The general predominance of magnesia over lime in the analyses and the small amount of lime show that calcic silicates have been largely destroyed during the process of erosion and the deposition of the sediments. The presence of carbon in black slates and of ferric iron in red slates is to be noted. Analysis V, of a slate consisting chiefly of sericite and partly of potassium feldspar, is of interest because it indicates a raw material source of potash.

Varieties. Roofing slates are compact, fissile varieties that can be split into thin, smooth plates. Slates of all different colors are used, but the most common color is a dark gray. Some slates after being taken from the quarry fade on continued exposure; the likelihood of

TABLE 4

Analyses of Slates

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	C	XyO[1]	Total
I...	59.7	17.0	0.5	4.9	3.2	1.3	1.4	3.8	4.1	0.5	3.8	100.2
II...	67.6	13.2	5.4	1.2	3.2	0.1	0.7	4.5	3.3	...	0.7	99.9
III...	59.8	15.0	1.2	4.7	3.4	2.2	1.1	4.5	3.8	...	4.3	100.0
IV...	56.4	15.3	1.7	3.2	2.8	4.2	1.3	3.5	4.8	0.6	6.5	100.3
V...	56.7	19.3	5.6	1.9	1.9	0.0	0.5	8.8	3.8	...	1.3	99.8

[1] XyO represents small amounts of other oxides.

I. Black roofing slate, Benson, Vermont (XyO = TiO_2, P_2O_5, CO_2, FeS_2, etc.).
II. Red roofing slate, Washington Co., New York.
III. Green roofing slate, Pawlet, Vermont (CO_2 = 3.0).
IV. Black roofing slate, Slatington, Pennsylvania (CO_2 = 3.7, FeS_2 = 1.7).
V. Slate, Cartersville, Georgia (Georgia Geologic Survey).

fading can only be determined by practical trial. Slates in which pyrite and marcasite are present in any notable quantity are commercially less valuable, as the sulfides, especially the marcasite, are likely to alter on exposure and give rise to rusty stains. Slates used for making blackboards are the blackest and most minutely grained kinds. *Calcareous* slates are those that contain much calcite, which may be as much as 30 per cent of the whole; they represent slates that have been formed from original marls.

Cleavage of slates and its origin. The cause of slaty cleavage has been the subject of much geologic, experimental, and mathematical investigation. Lately X-ray and other techniques have been employed to study slates. From the earliest work it became clear that slaty cleavage is the result of great pressure acting on the material. When clay sediments are subjected to intense pressure, the flaky particles (sericite, chlorite, and kaolinite) in them tend to rotate and thereby to set their broad sides perpendicular to the direction of pressure. The parallel arrangement of the flaky minerals thus induced gives the rock a closely spaced system of parallel planes, along which it splits more readily than along any others. The flaky minerals have an excellent cleavage parallel to their flat directions, and this property combined with the parallel arrangement of the flakes determines the capacity of the rock to split into thin smooth plates. Slaty cleavage is thus largely determined by the cleavage of the flaky minerals in parallel arrangement. Not all the minerals whose parallel arrangement and cleavage determine the slaty cleavage were necessarily original constituents of the clay; many of them, micas for example, have grown during metamorphism of the clay or shale to slate.

The planes of slaty cleavage do not necessarily have a definte relation to those of original bedding, Actually, as can be seen in the field in folded series of slates, the cleavage planes cut across the bedding at various angles, and in the arches of anticlines and in the troughs of synclines they cut across the bedding at right angles. The cleavage planes themselves, however, are strictly parallel to one another, as can be seen in the diagram in Fig. 72.

Besides having a cleavage, slates are intersected by cross joints, which may be so closely spaced as to divide them into small blocks and make them useless for technical purposes. The joints generally form systems intersecting at definite angles. In some districts slates have been affected by later folding and show puckering and crumpling on their cleavage surfaces in consequence of this later deformation.

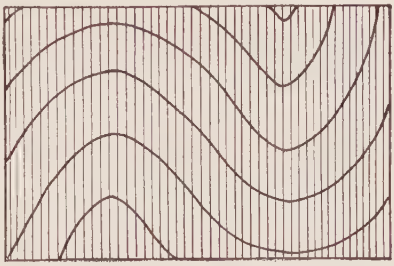


FIG. 72. Slaty cleavage cutting across folded beds.

Occurrence. Slates are common in metamorphic regions, representing belts of mild metamorphic intensity. They range in age from Precambrian to Cenozoic. They are chiefly Paleozoic in eastern North America, where they are abundant in Maine, Vermont, Pennsylvania, and Georgia. They occur also in the Lake Superior region, in the older ranges of the Rocky Mountains, and in many other parts of the world.

Phyllite. Closely connected with slate by intermediate varieties is a group of rocks called *phyllite*. Phyllites differ from slates in three particulars: (1) they are coarser grained than slates—they are microcrystalline; (2) they have a glossy luster, which slates do not have; and (3) many of them are porphyroblastic, containing conspicuous crystals of biotite, garnet, staurolite, and other minerals. The mica is the fine, scaly, silky variety of muscovite called *sericite*. Quartz is the chief constituent, and can generally be seen on the cross fracture. The rocks that in this country have been called "hydromica schists" are in large part such phyllites.

Their color ranges from silvery white, usually tinged with reddish, yellowish, or greenish tones, to dark colored or black, from pigments like those of slate. Some phyllites contain porphyroblasts of biotite, garnet, and staurolite.

As phyllites become coarser in grain, they grade into mica schists.

Consequently the divisional line between phyllite and mica schist is arbitrary.

Phyllites are of diverse origins. Some phyllites represent sedimentary material that has been metamorphosed, like the slates, but has attained a more advanced degree of reconstitution than the slates. Other phyllites represent felsites or felsite tuffs that were subjected to energetic metamorphism by strong shearing stress under heavy pressure. The feldspars have been largely, if not entirely, converted into sericite, and a thin schistose or slaty cleavage has been imposed upon the rocks. Some phyllites are so rich in sericite as to seem to be wholly composed of this silky mica; they are called *sericite phyllites*. Chemical analyses of these sericite phyllites show them to have a composition like that of many felsites and felsite tuffs.

TALC SCHIST

Talc schist is a markedly schistose rock in which talc predominates. The talc is present as flakes ranging from fine scales in some rocks to coarse foliated aggregates in others. Minerals also occurring in different varieties of talc schist are: magnetite and chromite in black specks and grains; amphibole, in white or green prisms; enstatite; olivine; and chlorite. Quartz, dolomite, and magnesite are uncommon.

The color is usually light: white to pale green, or yellowish, or gray; rarely it is dark gray or greenish. The rock is soft, and the talc gives it a greasy feeling and a pearly or tallowy appearance on cleavage surfaces. In addition to its micaceous habit and soft greasy feel, talc is easily recognized by its infusibility before the blowpipe and its insolubility in acids. Some talc schists are thinly schistose; others are more thickly schistose. If there is almost no foliation, the rock is nearly massive, is compact, has a waxlike aspect, and approaches soapstone in characters. Chemically, talc schists consist mostly of silica and

TABLE 5

Analyses of Talc Schists

	SiO_2	Al_2O_3	Fe_2O_3 + FeO		MgO	CaO	Na_2O + K_2O		H_2O	Total
I...	58.7	9.3	4.4	...	22.8	0.9			4.1	100.2
II...	53.3	4.4	5.8	1.0	22.9	1.5	1.5		2.6	100.0
III...	53.6	3.6	5.5		23.0	11.9	0.4	trace	2.5	101.6

I. Talc schist, Falun, Sweden.
II. Talc schist, Zöptau, Moravia, Austria.
III. Talc schist, Gorgona Island, Tuscany, Italy, containing 1.1 per cent Cr_2O_3.

magnesia with a small amount of water, about 3 or 4 per cent. The composition is quite similar to that of peridotites, as can be seen by referring to their analyses.

Talc schists represent in the main material that was of igneous origin -peridotite or pyroxenite; less commonly the material was of sedimentary origin—dolomitic marls. It may not be possible from field work and inspection of specimens alone, unless aided by chemical analyses and microscopical study, to decide from what rock a given talc schist is derived. Sometimes it is impossible to determine the origin by any means. The presence of chromium, either as chromite or as secondary minerals derived from it, such as kämmererite (chromiferous chlorite) or fuchsite (a variety of muscovite that is green from chromium), indicates derivation from ultramafic igneous rocks, and on the other hand the presence of much quartz and dolomite mingled with the talc, producing the variety of talc schist called *listwänite*, indicates sedimentary derivation.

Talc schists do not make extensive formations as do gneiss, mica schist, and slates but are intercalated chiefly as lenticular masses in low-rank metamorphic rocks and are not very common. They show transitions in places into other rocks, such as serpentinite, chlorite schist, dolomite marble, and quartzite. Such transitions, or the lack of them, give useful hints as to the origin of particular talc schists.

Commercial talc deposits appear to have been formed by the action of siliceous emanations from intrusive granite acting on dolomite. The Saint Lawrence County deposits in New York have resulted from the alteration of tremolite rocks, which had earlier been formed by contact-metamorphic action of granite intrusive into dolomite.

CHLORITE SCHIST AND PHYLLITE

Chlorite schist and phyllite, as their names imply, contain chlorite as their dominant mineral. The chlorite generally occurs as minutely scaly aggregates; if the individual scales are too small to be perceived by the unaided eye the rock is called a *chlorite phyllite*; if it is made up of coarser flakes it is called a *chlorite schist*. Chlorite phyllites and schists that are true, that is, consisting largely or wholly of chlorite, are poorly foliated and may in fact resemble serpentinites. The only obvious physical difference between such chlorite rocks and serpentinites is that they are softer, in fact, they can be scratched with the fingernail. The color ranges through shades of green, from yellow green to dark green. Other minerals may accompany the chlorite, chiefly such low-rank metamorphic minerals as sericite, albite, epidote, and actinolite. Abundance of epidote, for example, gives rise to the variety *epidote-chlorite schist.* The chemical composition of the chlorite foliates is highly varied: it indicates that they have been derived from rocks of different kinds, as indicated by the analyses in Table 6. Analyses I and II have a composition like that of gabbros, dolerites, and basalts, these being merely texturally different products of gabbroic magmas. These rocks are rich in pyroxene, which under shearing stress is converted into scaly chlorite. Analysis III suggests that this chlorite schist was derived from a peridotite; because of the high content of magnesium oxide, it was probably largely derived from olivine.

TABLE 6

ANALYSES OF CHLORITE SCHISTS

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	TiO_2	XyO[1]	Total
I...	49.2	15.1	12.9	...	5.2	10.6	3.6	1.5	1.9	...	...	100.0
II...	47.8	16.5	4.2	7.4	6.2	7.0	3.2	0.8	4.0	2.3	0.9	100.3
III...	31.5	5.4	10.2	...	41.5	...	...	...	9.3	...	...	97.9

[1] XyO represents small amounts of other oxides.

I. Chlorite schist, east of Rotön, Sweden.
II. Chlorite-epidote schist, Montezuma, Mitchell Co., North Carolina.
III. Chlorite schist, Pfitschertal, Tyrol.

True chlorite phyllites and schists are not particularly abundant rocks. They occur, for example, in the Mother Lode belt of California, where they were formed as results of the shearing of pyroxenite. However, chlorite-bearing phyllites, in which the chlorite is associated with sericite, quartz, and albite, are abundant and widely distributed. More rocks probably are called chlorite schist than would strictly qualify for this name.

Greenstone. Under the description of gabbro and of dolerite it was mentioned that these basic igneous rocks are transformed by the processes of metamorphism, depending on the physical conditions prevailing, into hornblende schists or into "greenstone" or "greenstone schist." During this metamorphism the original ferromagnesian minerals or the hornblende produced from them are largely changed into chlorite, which gives the resulting rock a green color. Such rocks are called greenstones if massive, and greenstone schists if schistose. Metamorphic alteration of the hornblende in diorites to chlorite also produces greenstones.

Transitions from dolerite and basalt into greenstone and greenstone schists are abundant in many regions. Lavas are likely to be transformed during mild metamorphism into greenstones, whereas tuffs and breccias intercalated between the lava flows are likely to be changed into greenstone schists. A greenstone schist from the Menominee River, Michigan, which is known to be an altered dolerite, has the following composition.

SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	CO_2	Total
44.5	16.4	5.1	5.5	7.5	7.9	2.6	0.5	5.0	5.4	100.4

It is composed of chlorite, with some feldspar, quartz, and calcite. This analysis should be compared with the analyses of gabbro which represent the gabbro-dolerite group.

Greenstones range from pale gray green through yellowish green to dark green. The color depends on the proportion of chlorite and epidote to other minerals. Greenstones are generally too fine grained for the megascopic determination of the individual mineral particles. If the original dolerite or basalt was amygdaloidal, the amygdules are likely to be retained, and the metamorphic rock is filled with little balls of calcite or quartz. Where the rock was strongly sheared, the amygdules have disappeared, but they are still represented in the resulting schist by ovoid spots differing in color and mineral composition from the main mass. Relicts of amygdaloidal structure are good proof of the igneous nature of the original rock.

Soapstone, or steatite. Closely connected with talc and chlorite schists is soapstone, or steatite, a massive rock, generally gray or green, less commonly a dark color; the lighter-colored rock has a silvery or shimmering fracture surface. Soapstone is very soft, is easily cut or worked, is without cleavage or grain, and resists well the action of heat and acids. For these reasons it was extensively used in prehistoric times in making pots and other vessels, and is now employed in table tops, sinks, laundry tubs, and other interior fittings where its chemically resistant qualities render it valuable. It is usually a variable mixture of interwoven scales of talc and chlorite and other minerals; carbonates may be present. The better grades are nearly pure talc rocks. The minerals are in general too fine for megascopic determination. Soapstone is associated with talc and chlorite rocks, and locally with serpentine, in areas of metamorphic rocks in various parts of the world.

HORNBLENDE SCHIST, AMPHIBOLITE, AND AMPHIBOLITE SCHIST

The amphibolites are a large group of metamorphic rocks, distinguished by the fact that they consist partly or largely of amphibole and are more or less schistose. There are a number of varieties in the group, depending on the kind of amphibole present and on the minerals associated with it, so that a general description covering all amphibolites cannot be given. It is best therefore to describe the most common kinds first, and then to mention briefly some of the less common varieties.

Hornblende schists carry quartz along with the amphibole, whereas amphibolite and amphibolite schists contain plagioclase feldspar instead of quartz; however, this distinction is generally difficult or impossible to make megascopically.

Amphibolites range in color from green to black; the green is of varied tones: clear light green, gray green, yellowish green, dark green, or greenish black; the darker colors are more common. The color is determined by the particular amphibole present, though in some amphibolites it is influenced by admixed chlorite. The grain size of the rocks ranges from fine to coarse. If it is coarse, the amphibole, which is generally present as slender prisms or blades and rarely as grains, is easily recognized by its form and its bright, good cleavage. The prisms may be an inch or more long and be as slender as match sticks; but in the finer-grained types they are tiny needle- or hairlike prisms that can be seen only by careful inspection with a good lens. The prisms as a rule lie in the surface of schistosity, and they may have a parallel arrangement within this surface. The parallel arrangement gives the rock, especially the finer-grained varieties having needle-like prisms, a rather characteristic shimmering or silky luster on its foliation surfaces.

Amphibolites are rather hard rocks, not easily scratched by the knife. The more schistose kinds are brittle, but the more massive are tough and difficult to break. They are heavy, the specific gravity ranging from 3.0 to 3.4.

Besides hornblende, other minerals occur in varying kinds and quantities; prominent are quartz, feldspar, and mica. The quartz and feldspar in grains are best seen with the lens on the cross fracture, but they may be too fine and too much masked by the hornblende to be seen. The quartz forms little lenses or masses; if it has been secondarily deposited from solutions it fills fractures in the shape of gash veinlets, as in other metamorphic rocks. The mica can generally be seen on the surface of chief fracture; both biotite and muscovite occur, possibly to such an extent as to produce transitions to mica schist.

Other minerals megascopically recognizable are iron ore, garnet in small dark-red crystals, chlorite, and calcite; the calcite may occur as veinlets, as quartz does. Pyroxene, epidote, and other minerals occur, but partly because of the fineness of grain and partly because they resemble hornblende, they are difficult or impossible to detect and identify without microscopic study.

The chemical composition of amphibolites shows, in agreement with facts to be presently mentioned, that the origin of these rocks is varied. In Table 7, analyses I and II are much like that of the gabbro-basalt magmas, as can be seen by comparison with analysis V. Specimen III, which is almost lacking in aluminum oxide, has the general composition of the peridotite-pyroxenite group of igneous rocks and may be compared with analysis III of that group. The presence of chromic oxide in specimen III is also significant of the derivation of this amphibolite from peridotite. On the other hand specimen IV is thought on geologic grounds to be derived from an impure dolomite containing much clay, and this supposition is rendered probable by the fact that the abundant alumina is accompanied by almost no alkalies, a feature not seen in igneous rocks.

TABLE 7

ANALYSES OF AMPHIBOLITES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	XyO[1]	Total
I...	49.9	15.5	3.0	8.0	7.8	8.9	3.3	0.7	1.5	1.7	100.3
II...	50.4	13.3	6.3	9.3	5.6	7.9	2.1	1.1	1.7	2.0	99.7
III...	55.0	2.9	0.8	6.3	21.0	11.5	0.3	0.2	1.0	1.5	100.5
IV...	55.6	16.3	1.2	7.2	5.6	9.2	0.9	0.2	3.1	1.0	100.3
V...	52.4	13.6	2.7	9.8	5.5	10.0	2.3	0.4	1.7	1.6	100.0

[1] XyO represents small amounts of other oxides.

I. Thin schistose amphibolite, Whitman's Ferry, Sunderland, Massachusetts.
II. Amphibolite, Crystal Falls district, Michigan.
III. Grass-green amphibolite, Chiavenna (XyO = Cr_2O_3).
IV. Amphibolite, Goshen, Massachusetts.
V. Olivine basalt, main lava flow, Pine Hill, South Britain, Connecticut.

Origin of amphibolites. It has just been shown that some amphibolites correspond in composition to igneous rocks, but others do not, and this fact agrees with the results of geologic investigation in the field. For in some places we find amphibolites under conditions that strongly suggest that they were derived from igneous rocks, and in other places either such evidence is wanting or the contrary is indicated. Chemical criteria and the use of the microscope on thin sections in which the textures and associated minerals can be seen confirm the conclusion that amphibolites are of dual origin, formed from both igneous and sedimentary rocks.

In the description of gabbro and dolerite it was mentioned, under alteration, that these rocks become converted by shearing into hornblende schist or amphibolite. Gradual transitions are found from unaltered gabbro and dolerite into wholly metamorphic amphibolite, without breaks in continuity. Thus, as the feldspathic igneous rocks give rise to gneisses, so the ferromagnesian igneous rocks, especially the pyroxenic ones, give rise mainly to hornblende schists, and also to talc schists, chlorite schists, and serpentinite.

If subjected to metamorphism, sedimentary beds of mixed composition, such as limestones containing sand, clay, and more or less iron hydroxide, or marls of somewhat similar nature, might under suitable conditions be converted largely into hornblende, mixed with other minerals. The volatile constituents—water and carbon dioxide—are largely driven out during metamorphism; the bases—lime, iron, magnesia, and alumina—combine with the silica to form silicates, of which hornblende is the chief and the determinant one in the resulting rock. Thus amphibolites may result from the metamorphism of argillaceous dolomite, as described later under marble.

Varieties of amphibolite. In the midst of gneisses and mica schists amphibolite may be massive. The prisms and grains of hornblende, instead of being arranged in parallel position and thus producing a schistose foliation, are interwoven in random arrangement, and consequently foliation is wanting. In such massive amphibolites the hornblende is likely to be accompanied by feldspar.

Glaucophane schist is a variety of amphibole schist in which the distinctive amphibole is the blue sodium-bearing species, glaucophane, and for this reason the schist is colored rich blue instead of green. Other minerals may be present, depending on the origin, such as quartz, epidote, pyroxene, chlorite, and garnet. Some glaucophane schists are coarse grained and these varietal minerals in them can be perceived; others are fine grained, slaty-blue or blue-gray rocks. Although glaucophane schists are comparatively rare, they are widely distributed in the Coast Ranges of California and Oregon and in other parts of the world.

Occurrence of amphibolites. Amphibolites form layers and masses in other metamorphic rocks of corresponding metamorphic rank, especially in gneisses and mica schists. They occur also as extensive independent formations. Some occur in gneisses in long bands in such a manner as to suggest that they are metamorphosed dikes of dolerite. In size the masses range between wide bounds, from a foot to thousands of feet in diameter. In some places, they make up in frequency of occurrence for what they lack in size. The manner in which they are interlayered in places with other metamorphic rocks suggests that some were formed from sills of basic igneous rock and some from interbedded argillaceous dolomites and layers of basalt tuff, but in general these inferences can be made more probable only by supporting chemical and microscopical investigations.

Amphibolites are extremely common rocks in metamorphic regions.

MARBLE AND CARBONATE-SILICATE ROCKS

Marble is the metamorphic equivalent of limestone and dolomite. It is distinguished from these rocks by its livelier sparkle and purer colors. Pure calcite marbles contain only calcium carbonate, and pure dolomite marbles contain only dolomite. If the limestone and dolomite from which marbles are derived contained impurities, additional kinds of marble were formed, differing as to the associated minerals they contain. The different kinds are described separately after a general description.

General properties. Marble is a crystalline granular rock composed of calcite grains. The grain size ranges from fine, in which individual grains cannot be distinguished, to coarse; in the coarsest varieties the cleavage surfaces of individual grains are half an inch or more wide, but these are uncommon. The fracture surface of the finest-grained marbles used for statuary has a soft shimmering luster, while the appearance of coarser kinds is like that of cube sugar.

The normal color is white, snow white in the best statuary marble such as the famous Carrara marble; but many marbles are more or less colored by pigmentary substances, chiefly carbonaceous matter and iron oxides. They are consequently gray, black, yellow, or red. Although the color may be uniform, it is commonly spotted, blotched, clouded, or veined, producing the effect known as "marbled." The hardness is that of calcite, 3; marble is therefore readily scratched or cut by the knife, a ready means of distinction from quartzite or sandstone, which may resemble it. It dissolves easily in weak acids. Unless the grain is too fine, the good rhombohedral cleavage of the calcite grains can be easily seen with a lens.

The chemical composition of a perfectly pure marble is that of calcite ($CaO = 56$ per cent; $CO_2 = 44$ per cent), but small quantities of magnesia, alumina, iron, and silica are generally present, coming from traces of sand, clay, and dolomite mixed with it. If these extraneous constituents are abundant, impure marbles, described in a later paragraph, are produced

Unlike most metamorphic rocks, marble, if pure, is generally massive and devoid of foliation even where it is interbedded with schists and gneisses. If it contains impurities in the form of other minerals, it may have a foliation caused by their presence. The lack of foliation in marbles is an interesting matter. The explanation appears to be that calcite recrystallizes readily, far more readily than other rock-making minerals, and during this recrystallization whatever foliation had developed is obliterated. Some rare marbles, however, are markedly schistose. The well-foliated Talledega marble of Alabama is an example.

Varieties of marble. The varieties of marble technically valuable are chiefly distinguished on the basis of color. *Statuary marble* is the purest, whitest, and most homogeneous kind. *Architectural marbles* have the most uniform tones of color, while *ornamental marbles* are distinguished by striking effects of varied colors.

In the trade, marble means any calcium carbonate or dolomite rock that can be procured in large, firm blocks and can take a high polish. According to this usage many limestones are called marbles. The widely used "Tennessee marble" is a nonmetamorphic limestone of great purity occurring near Holston in east Tennessee; it consists chiefly of crinoid fragments in a groundmass of distintegrated bryozoa. The yellow, red, and black marbles, most of them veined and clouded, of Italy, Greece, and the East have long been distinguished by a host of names.

The varieties that are distinguished by the presence of some mineral in addition to the calcite are treated below in the section on carbonate-silicate rocks.

Occurrence of marble. The great deposits of marble from which the material used for structural purposes is taken occur in regions of metamorphic rocks associated with gneisses, schists, and phyllites. These marble deposits range in length within wide limits, from a few feet to many miles. Marble forms thick interbedded layers in the Precambrian Grenville rocks of Canada. It occurs in quantities in Vermont, Massachusetts, and Georgia where it is extensively exploited; in Colorado, and in other places in the West. The marbles of Greece and Italy attained celebrity from their use by the ancient Greeks and Romans in statuary and buildings. Marble occurs in many other places in the world.

Marble has also been produced from limestone by the contact-metamorphic action of intrusive igneous rock. The Yule marble in the heart of the Rocky Mountains in Colorado, the handsome marble used to make the tomb of the Unknown Soldier in Arlington, is said to have been formed in this way.

Calcium carbonate-silicate rocks. As was stated under the general properties of sedimentary rocks, all transitions occur between limestones and sandstones, between limestones and shales, and among the three combined. This means that the original calcareous deposits have had sand, clay, silt, and ferruginous material in variable amounts mixed with them. Chemically, it means that the calcium carbonate has silica, oxides of aluminum and iron, and small amounts of other oxides, such as magnesia, potash, and soda, mixed with it. Under the conditions of metamorphism the carbon dioxide of the calcium carbonate is liberated and is driven out; the lime (CaO) thus set free combines with the silica present, and thus silicates of calcium, of calcium and aluminum, or of calcium and iron, mixtures of these, or silicates containing other elements as well are formed. Furthermore, volatile substances, liquids and gases, such as water furnishing hydroxyl, fluorine, and boron, may enter the rock mass as emanations from intrusive magmas and thus, by adding new substances, produce other new minerals. The amount of silica present may suffice to replace the carbon dioxide completely and thus produce a rock composed entirely of silicates; it may not be sufficient. If it is not, the newly formed rock consists of a mixture of calcite and silicates. Thus all transitions may be found from pure marble, through varieties containing scattered crystals and sporadic masses of some mineral or minerals, into rocks wholly made up of one or, usually, more silicates. The reconstitution of the sedimentary rock of mixed composition is quite analogous to what has already been described in this book as the effect of contact metamorphism on impure limestones, and the chemical reactions that take place are the same as those there mentioned. The resulting rocks also are similar, with one difference however. In contact metamorphism the chief agent of metamorphism is heat, and shearing stress is not operative, but in regional metamorphism powerful shearing stresses under strong confining pressure have been at work. Consequently, the rocks formed by contact metamorphism are massive, having little or no foliation, whereas those formed by regional metamorphism characteristically have well-developed foliation.

Important minerals that thus occur in limestone are *wollastonite* ($CaSiO_3$), pyroxenes, especially diopside, $CaMg(SiO_3)_2$; garnet, as grossularite, $Ca_3Al_2(SiO_4)_3$; amphiboles, especially tremolite, $H_2Ca_2Mg_5Si_8O_{24}$; feldspar, especially anorthite, $CaAl_2Si_2O_8$; vesuvianite; and epidote. All these minerals are high in calcium, as is to be expected, because they were formed in a high-calcium environment. Many other minerals occur, such as phlogopite, graphite, magnetite, spinel, and apatite, and were formed from the less abundant impurities in the original limestones.

It is clear that, depending on the original chemical composition, a great variety of calcium carbonate-silicate rocks exist; only a few of the most important will be mentioned.

Wollastonite marble. Some marbles contain crystals of wollastonite ($CaSiO_3$), and in some the wollastonite is so abundant that it composes practically the entire rock. Wollastonite rock is white and is generally massive. It is widely distributed as interbeds in crystalline schists, but is not abundant.

Garnet rock, or garnetite. Garnetite is an interlocking aggregate of grains of garnet, generally accompanied by other minerals in minor amount. Since garnetites have resulted from the metamorphism of limestones, calcite is the mineral most likely to be present. If calcite occurs, the garnet tends to have more or less well-developed crystal forms. From outcrops of garnetite the calcite has generally been leached out and consequently such rock is porous. Garnetites tend to be yellowish or reddish brown.

Garnetites composed of andradite, $Ca_3Fe_2^{+++}(SiO_4)_3$, are common as metasomatic products in contact-metamorphic aureoles. Such andradite rocks and associated andradite-pyroxene rocks are called *tactites*. As many tactites are ore bearing, containing ores of copper, tungsten, and other metals, they are of much economic interest.

Garnetites formed during regional metamorphism are much less common. A notable example of a garnetite of regional-metamorphic origin is the Loch Tay limestone, in Scotland; it consists chiefly of grossularite, $Ca_3Al_2(SiO_4)_3$, enclosing long prisms of vesuvianite, another high-calcium mineral. It is the product of regional metamorphism of medium intensity.

Pyroxene rock. The chief constituent is pyroxene, of which the variety diopside is prominent. Other minerals, such as quartz or calcite, may occur. The rock is white, greenish, or dark green, massive or schistose. Under the head of metamorphic pyroxene rocks there may be mentioned jade, which, although extremely rare, is of great interest from its ethnological and artistic importance. *Jade* is a fine-grained, and usally compact, aggregate of grains and fibers of the sodium pyroxene, jadeite ($NaAlSi_2O_6$). Some jade is snow white, resembling marble, but characteristically it is greenish (or with a violet shade) to dark green. The greenish colors are also clouded, veined, or specked through the white. When polished it has a soft, somewhat greasy luster. The extraordinary toughness of jade is one of its most marked characters, and on this account jade was greatly prized in the early history of mankind, before the discovery of metals, for the manufacture of weapons and implements, as is shown by its distribution in these forms and in unworked pieces over the world. It has long been greatly valued by the Chinese, who have devoted the most laborious work to fashioning it into objects for personal adornment and use. Many of these objects, such as vases and bowls, are carved with wonderful skill and taste and are greatly prized for their artistic value. The rock is known to occur in place only at Tammaw in upper Burma, where it is associated with serpentinite and glaucophane schist; in the Kuenlun Mountains, Turkestan; in Piedmont and Liguria, Italy; and in Wyoming. Its origin is uncertain. A green hornblende rock called *nephrite*, from Siberia and New Zealand, has similar properties and uses and is frequently mistaken for jade.

Dolomite marble, magnesium silicate rocks. Just as calcite marble is related to limestone, so dolomite marble is related to dolomite rock. Like calcite marble, dolomite marble is one end of a series of metamorphic rocks, which, beginning with a pure carbonate, becomes a mixture of carbonates and silicates and ends in pure silicate rocks. The causes and processes are identical with those described under calcite marble, but in dolomites the presence of magnesium causes the formation of silicate minerals in which magnesium is either the only metal, as in the magnesian olivine (Mg_2SiO_4), or is very abundant, as in tremolite [$Ca_2Mg_5Si_8O_{22}(OH)_2$]. The magnesian silicates thus formed by metamorphism are anhydrous, or nearly so; however, they may be secondarily converted into serpentine ($H_4Mg_3Si_2O_9$) or, less commonly, into talc [$H_2Mg_3(SiO_3)_4$]. The more important magnesium silicates taking part in the series are olivine, diopside, tremolite, and phlogopite. More rocks are formed by these mixtures than can be mentioned here.

Dolomitic marbles containing variable amounts of phlogopite, tremolite, and pyroxene, with others, such as magnetite, spinel, apatite, and graphite, derived from original impurities, occur rather commonly in metamorphic areas.

Ophicalcite is a mixture of white calcite and green serpentine, the latter in veins or spots or clouded through the rock. The "verde antique" marble of the ancients, used for ornamental purposes, appears to have been in part a variety of ophicalcite.

Amphibolites. Many of the amphibole schists, amphibolites, and amphibolite schists, previously described, are products of the transformation of impure dolomites into metamorphic rocks. This subject has already been discussed, but it is mentioned again because amphibole-rich rocks made in this way are important members in the calcium and magnesium series of carbonate-silicate rocks described above.

SERPENTINITE

General properties. In describing the mineral serpentine the chemically pure substance was considered, but serpentine as a rock, for which the name serpentinite is coming into use, has more or less of other minerals mixed with it. Serpentinites are generally compact, of a dull to waxy luster, and of a smooth to splintery fracture. If tolerably pure, they are soft and can be cut by the knife, but some are saturated with secondary silica, which makes them much harder. The general color is green, characteristically a yellowish green; but yellow, yellow brown, reddish brown, and dark green to black are common. On smooth surfaces serpentinite has a somewhat greasy feel, recalling talc schist, from which, however, it is readily distinguished by its superior hardness. Talc leaves its mark on cloth, and serpentine does not. The yellow-green color resembles that of epidote rocks, but the inferior hardness of serpentine serves to distinguish it from epidote.

Associated minerals. Other minerals that commonly occur in serpentinite are remains of the magnesium silicates from which it was formed: olivine, pyroxene, and hornblende. Metallic-looking specks and crystals of magnetite and chromite are common.

Serpentinite is the source of the platinum in the placers of the Ural Mountains, and of the nickel ores in New Caledonia and other places. Serpentinite generally contains other secondary minerals: chlorite (the purple-red variety kämmererite containing chromium); talc; and the magnesium carbonates magnesite ($MgCO_3$), and breunnerite ($Mg, FeCO_3$). Serpentinites are usually massive, but some are schistose (serpentine schist). At some places they are seamed by veins of the finely fibrous variety called chrysotile, which is generally called asbestos.

Chemical composition. The chemical composition of serpentinite approaches that of the pure mineral, but it generally differs somewhat because other minerals are present in minor amount, as can be seen from the analyses in Table 8. The presence of nickel and chromium in small quantities is common, because most serpentinites have been derived from peridotites and pyroxenites.

Origin. Serpentine rocks are secondary in origin, being formed by the hydration of previously existent rocks consisting wholly or chiefly of magnesium silicates. Their origin is therefore twofold: some have been formed from igneous rocks, such as peridotite and pyroxenite,

TABLE 8

ANALYSES OF SERPENTINITES

	SiO_2	Al_2O_3	Cr_2O_3	Fe_2O_3	FeO	NiO[1]	MgO	CaO	H_2O+	CO_2	XyO[2]	Total
I.	40.7	1.0	0.3	6.6	1.2	0.2	36.4	0.1	12.4	0.2	1.0	100.1
II.	41.7	0.8	0.4	4.1	2.1	0.1	38.5	0.4	11.0	0.0	0.6	99.7
III.	39.8	1.4	0.3	10.1	...	1.0	33.7	...	13.3	...	...	99.6
IV.	43.3	...	...	...	...	...	43.7	...	13.0	...	...	100.0

[1] The NiO as reported generally contains a minor amount of CoO.

[2] XyO represents small amounts of other oxides.

I. Serpentinite, derived from dunite, Cornwall, England. Felted mass of fibrous and scaly serpentine, with sporadic bastite pseudomorphs after orthopyroxene. Analyst, E. G. Radley.

II. Serpentinite, derived from pyroxenite; "almost black"; Mount Diablo, California. Analyst, W. H. Melville.

III. Serpentinite, from which the lateritic iron ore of the Mayari district, Cuba, was derived.

IV. Theoretical composition of pure serpentine, $Mg_3Si_2O_5(OH)_4$.

and others from metamorphic rocks—amphibolites or hornblende schists, some of which have been formed by the metamorphism of impure dolomites. Thus the origin of the material may be igneous or sedimentary, but, whereas igneous rocks can be altered directly into serpentine, sedimentary ones must first pass through an intermediate stage, namely the metamorphic stage (hornblende schists, etc.), and later may become converted to serpentine. No formula can be given for determining the origin of a serpentinite. However, the geologic mode of occurrence of the serpentinite and its relation to other rock masses is often a help, and if nickel and chromium—common substances in ultramafic igneous but not in sedimentary rocks—are shown to be present in a serpentinite their presence is highly significant, amounting practically to proof that the serpentinite in question was derived from an igneous rock.

Occurrence. Serpentinite is a common rock, occurring wherever masses of peridotite and other ultramafic rocks are found. As layers and lenticular masses, it is common in metamorphic regions from the alteration of highly magnesian igneous and metamorphic rocks.

Alteration. Uses. Serpentinite strongly resists the action of the weathering agencies at the Earth's surface, but it eventually breaks down into a mixture of carbonates and silica, mixed with ferruginous matter. The soils thus formed lack alkalies and lime and are infertile, and little or no vegetation grows on them.

In Cuba, lateritic weathering of serpentinite has produced enormous valuable iron-ore deposits, whose reserves exceed 1000 million tons.

Because of its beautiful coloring, serpentinite has been extensively quarried as an ornamental stone, being used for interior decoration much as highly colored marbles are. In some places, notably at Thetford, Quebec, veinlets of fibrous chrysotile ramify through serpentinite and are mined for use as asbestos. The importance of serpentinite as the host rock of ores of nickel and chromium has already been mentioned.

IRON OXIDE ROCKS

Itabirite. Itabirite is composed chiefly of micaceous hematite and quartz. The micaceous hematite is in very thin tablets or leaves of irregular outline, and the quartz is in aggregates of grains. Itabirite much resembles mica schist, and if one were to imagine the mica of such a schist replaced by a substance of mica-like thinness and the metallic luster of polished iron, one would have a good idea of how itabirite looks. Micaceous hematite is indeed of not infrequent occurrence in mica schists, and by its increase transition forms to itabirite are produced. Also, just as the relative quantities of quartz and mica vary in different layers of mica schist, so do the micaceous hematite and quartz vary in itabirite; thus there are some layers poor in quartz, and others rich in it, of greatly differing thicknesses. Itabirite is generally granular to fine granular, very schistose, and dark in color on the cross fracture; it exhibits on the chief fracture the shining steel-like luster of the specular hematite. In some specimens the amount of hematite is so great as practically to conceal the quartz. Itabirite occupies extensive areas near Itabira in Minas Gerais, Brazil, whence came the name. It was probably formed by the metamorphism of siliceous sediments rich in ferruginous matter.

Jaspilite. Banded rocks that consist of alternate layers of red chert and hematite are called jaspilite. They occur abundantly in the Precambrian iron formations of the Lake Superior region; identical material occurs also in the Precambrian of India. Jaspilite is metamorphosed ferruginous chert. See page 274.

Magnetite rock. Magnetite rock is a compact to granular aggregate of grains of magnetite, dark colored to black, and heavy. Hematite is commonly mixed with it, as are various other minerals, such as ilmenite, pyrite, quartz, calcite, and garnet, according to the mode of occurrence. Magnetite rock has originated in several ways. It occurs in masses inclosed in, or associated with, igneous rocks and is then regarded as a differentiated facies of such rocks. The associated minerals vary with the kind of rock: nepheline and augite occur with nepheline syenite (Arkansas, Brazil, Sweden), for example, and olivine, pyroxene, and calcic feldspar occur with gabbros (Adirondacks, Sweden, Canada). Magnetite rock occurs also as a contact-metamorphic product where igneous rocks have metamorphosed beds of limonite, siderite, and other iron-bearing minerals. It occurs in areas of regional metamorphism as layers and lenses in the midst of gneisses and schists, associated with metamorphosed limestones and dolomites. It then contains calcite and dolomite, as well as the more common silicate minerals described as associates of marble, such as garnet, pyroxene, and amphibole. The magnetite rock is probably the result of the metamorphism of beds of impure limonite, clay ironstone, etc.

Deposits of magnetite rock occur in many places in the United States and elsewhere and are of economic importance as sources of iron ore. The titaniferous magnetites, which are differentiates of gabbro and anorthosite, were until 1944 generally useless because they contain ilmenite, which prevented their being profitably smelted. By 1944 titanium had become sufficiently valuable so that extraction of the ilmenite as a source of titanium became profitable, and the magnetite was recovered as a by-product and utilized as iron ore. Mining and milling of titaniferous magnetite was successfully begun in 1944 at Lake Sanford in the Adirondack Mountains in New York.

TABLE OF PRINCIPAL METAMORPHIC ROCKS

Table 9 is intended to give an over-all picture of metamorphism and the resulting metamorphic rocks. Details are omitted, only the more important metamorphites being shown.

The source material is shown in column I, the metamorphic rocks produced from the source material are shown in column II, and the chief constituents, including the index minerals or most distinctive minerals, are shown in column III. For example: granite, its corresponding extrusive equivalent (rhyolite), and arkose, all three being rocks of essentially the same chemical composition, tend to yield the same metamorphic products. The resemblance between the products derived from granite, rhyolite, and arkose becomes greater and greater as the rank of metamorphism becomes higher. The products of metamorphism are shown in column II as falling into two groups: hornfelses, which are massive, and foliates, which are more or less slaty, schistose, or gneissose. The foliates are divided according to degree of metamorphism into low-rank, medium-rank, and high-rank rocks. The minerals indicative of these ranks are shown in column III.

TABLE 9

Principal Metamorphic Rocks

I. Source Material	II. Corresponding Metamorphic Rocks	III. Chief Constituents (Index and distinctive minerals are italicized)
Granite and corresponding extrusive rocks Arkose	Hornfels	Of rare occurrence, not considered here
	Foliates	
	Low rank	
	Sericite phyllite Epigneiss	*Sericite,* quartz
	Medium rank	
	Muscovite schist and gneiss	Muscovite, quartz, potassium feldspar, biotite
	High rank	
	Katagneiss	Quartz, orthoclase, biotite, and minor *sillimanite, garnet, cordierite*
Shale	Hornfels	*Andalusite, cordierite, biotite,* quartz, feldspar
	Foliates	
	Low rank	
	Slate Phyllite	*Sericite, chlorite,* quartz
	Medium rank	
	Mica schist	*Biotite, muscovite,* quartz, feldspar
	High rank	
	Aluminum silicate gneiss	*Sillimanite, cordierite, almandite,* quartz, feldspar, biotite, muscovite
Sandstone	Quartzite (Metaquartzite)	Quartz, with subordinate amounts of minerals formed from the cement: *tremolite, epidote, diopside, wollastonite, biotite, muscovite, garnet, kyanite, sillimanite*
Gabbro and corresponding extrusive rocks Marls	Hornfels	*Plagioclase, pyroxene*
	Foliates	
	Low rank	
	Greenstone schists Greenschists	*Chlorite, epidote, actinolite, albite,* quartz, calcite
	Medium rank	
	Amphibolite	*Hornblende, pyroxene, plagioclase, garnet*
	High rank	
	Eclogite	*Pyropic garnet, pyroxene (omphacite)*
Peridotite and pyroxenite	Hornfels	Of rare occurrence, not considered here
	Foliates	
	Chlorite schist	*Chlorite*
	Talc schist	*Talc*
Limestone and dolomite (impure)	Calcic hornfels	*Diopside, tremolite, grossularite, et al.*
Limestone (pure)	Calcite marble	*Calcite*
Dolomite (pure)	Dolomite marble	*Dolomite*

CHAPTER XIV

DETERMINATION OF ROCKS

> A knowledge of rocks can never be gained from instructions given in books, but must be acquired by actual handling and study of the rocks themselves.—SIR ARCHIBALD GEIKIE.

The amount of difficulty met in determining and classifying rocks depends on how precisely we seek to classify them. The fine distinctions between rocks made by petrographers cannot of course be made in ordinary practice, unless we use petrographic methods—the microscope and chemical analyses. We cannot ordinarily employ such methods, and consequently we are limited to the means of examining rocks described in this book and to simple classifications and the small number of rock names that these classifications afford.

Rock characters used in determination. The characters used in determining rocks megascopically are of two kinds: *mineral* and *general*. If a rock is composed wholly or partly of mineral grains large enough to be distinctly seen with the unaided eye or with the lens and that can be handled and tested if necessary, we proceed to identify the rock by studying its minerals: the kinds, the relative abundance, and their relation to one another (rock fabric). In this way we determine the *mineral characters*. The individual minerals can be studied and tested according to the methods given in Chapter V. In the field, the simple tests of Table 1 (page 96) can be used. If the facilities of a laboratory are at hand, the more detailed Table 2 can be used. Even in the field, if a rock has already been determined to be igneous, sedimentary, or metamorphic, its more precise classification can usually be quickly made. Even if all the minerals in a rock cannot be identified, determination of one or more of them is as a rule helpful.

The *general characters* are the aggregate effects produced by the component mineral grains. They are *color*, *structure*,* *texture*, *hardness*, *fracture*, and *specific gravity*. Of these, specific gravity is the least easy to use, because it requires a special apparatus to determine it. The reaction of a rock with acids is also at times extremely useful

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* The difference between the use of "structure" and "texture" has been explained on page 129.

as a test, and should therefore be added to the list. The general characters are so helpful that they deserve separate mention in regard to their use in determining rocks.

Color. The color of a rock is the resultant of the colors of all the component mineral grains. Some conclusions can be drawn from the color of a rock; thus if the color is pure white or nearly so, iron-bearing minerals either are wanting in the rock or occur only in traces, and in general the rock is either sandstone, quartzite, marble, gypsum, or a nearly wholly feldspathic igneous rock, such as anorthosite, aplite, syenite, or felsite. Red, brown, and green colors indicate the presence of iron compounds. Black or stone gray also may indicate this, but in a sedimentary rock these colors may result from carbonaceous matter.

The "color of a rock" is the color it has on a fresh unweathered (newly fractured) surface. On a weathered surface the color may be vastly different.

Structure. If a rock has a pronounced structure, this structure may be of great assistance in determining the general class to which the rock belongs, especially if the geologic relations of the rock mass cannot be determined. Thus if a rock mass has a pronounced columnar structure or a highly vesicular or amygdaloidal structure, it is almost certainly of igneous origin. If it has a laminated or banded structure, it is probably sedimentary; however, banding cannot be relied on unreservedly, because some igneous rocks, especially lavas, have a banded structure as a result of flowage, and some metamorphic rocks have acquired one by shearing movement. Oölitic structure indicates a sedimentary rock. In general, structure must be considered in connection with texture and other properties.

Texture. Certain textures are of definite assistance in determining the major class to which a rock belongs. Thus glassy texture is definite proof of igneous origin. Porphyritic texture indicates the same thing, especially if the phenocrysts are quartz, or feldspar, or both, but this fabric is simulated in metamorphic rocks by the porphyroblastic fabric. Many metamorphic rocks contain prominent well-formed crystals set in a finer matrix, which simulate phenocrysts, as, for instance, garnet and staurolite, but in general these rocks have also a well-marked foliation, which helps to distinguish them. Sedimentary rocks do not have porphyritic fabric; a mere contrast in color produced by a few dark mineral grains scattered among many lighter ones must not be mistaken for porphyritic fabric.

Hardness. The hardness, readily determinable in a rough way in the field, is very useful in distinguishing between certain rocks. Thus many fine-grained compact sandstones (or quartzites), limestones, and aphanitic igneous rocks look much alike in specimens, but testing their hardness with the knife point at once distinguishes the limestones (carbonate rocks, soft) from the others mentioned (silicate, or silica rocks, hard). If a rock is not very coherent, the mere breaking down or crushing of its fabric should not be mistaken as actual scratching of its component minerals. If the rock itself is used to scratch with, care should be taken to test a number of corners, or edges, so that some single grain, harder than the average, may not produce a false impression of the average hardness.

Fracture. The fracture is less important than the foregoing characters, but is of some value. Most rocks that are firm and solid enough to have a distinct fracture break with a more or less rough, hackly fracture. Those that are fine grained or aphanitic and are composed largely or wholly of silica, such as felsites, quartzites, and chert, have a more or less distinct conchoidal fracture, and the surface may be splintery. Some fine-grained limestones also have a splintery fracture, which may even approach the conchoidal. Natural glasses, such as obsidian, have a beautiful conchoidal fracture; in them, in fact, conchoidal fracture reaches its acme of perfection.

Specific gravity. The specific gravity is of much more value in the determination of minerals than in that of rocks. It cannot, of course, be used in the field, as it requires definite apparatus to determine it, as described in Chapter III. Nevertheless, even in the field, rocks of low specific gravity, colloquially "light" rocks, can be roughly distinguished from "heavy" rocks by weighing them in the hand ("hefting them"). Rocks that are dark colored and very heavy are as a rule composed largely or wholly of iron-bearing minerals, and are likely to be of igneous or of metamorphic origin.

Treatment with acid. The test with acid is particularly useful in distinguishing carbonate from silicate rocks. The method of treatment is fully described in Chapter V. If necessary, almost any acid can be used, such as vinegar (acetic acid), or lemon juice. For field use a few crystals of citric acid can conveniently be carried and when needed dissolved in a little water; the test for effervescence can thus be readily made. The test for gelatinization, as described in Chapter V, also is useful in distinguishing nepheline syenite from other syenites and from granites and also in separating phonolite, the effusive representative of nepheline syenite, from the other felsites. It should be remembered that olivine, which, however, chiefly occurs in the dark ferromagnesian rocks (gabbros, peridotites, and basalts), also gelatinizes when subjected to the gelatinization test.

Determination in the field. The best method of determining the major class to which a rock belongs, that is, whether it is sedimentary, igneous, or metamorphic, is to study its characters in the place in which it occurs, and its relation to other rock masses. These features, and the larger ones of its structure, may be plainly apparent in the field, whereas a hand specimen may fail entirely to show them. The structure of a granite gneiss, for instance, may be conspicuous on the surface of a field exposure but quite inappreciable in a small specimen. It is not necessary to give here the characters and relations by which the major class can be determined; this is a geologic rather than a petrographic problem and has been sufficiently commented on in Chapters V, VIII, and X.

If the major class has been determined in the field, and the rock is coarse grained so that its constituent mineral grains can be seen and, if necessary, handled, Table 1 (page 96) of Chapter V can be consulted for their identification. Then by reference to the classification of the appropriate major class, the place of the rock in the classification can readily be determined.

Table for rock determination. Appended to this chapter is a table for determining the more important kinds of rocks. It is based essentially on that given by Geikie in his *Textbook of Geology*, but it has been considerably modified and extended to meet the needs of this book. As the tests that it demands are simple, consisting mainly of those relating to hardness and effervescence with acid, it can readily be used, even in the field. However, a table of this kind can be only quite general in character, and it applies only to rocks of well-defined types. Rocks grade into one another in so many ways, as has been described throughout this book, that not only the student, but even the experienced geologist, will sometimes be puzzled as to the proper naming of a particular specimen. If this limitation is borne in mind, however, the table will prove a useful aid in classifying the common rocks.

TABLE FOR DETERMINING THE COMMON ROCKS

The newly fractured surface of the unweathered rock shows one of the following features:

A. Glassy, wholly or partly. See *A* below.
B. Not glassy; dull or stony in appearance; homogeneous, without distinctive fabric; so fine grained that the individual grains cannot be seen and recognized. See *B*.
C. Distinctly granular to the unaided eye; the grains can be perceived and determined. See *C*.
D. Distinctly foliated. See *D*.
E. Clearly fragmental in composition. See *E*.

A. Wholly or partly glassy.

1. Wholly glass; solid; brilliant vitreous luster. *Obsidian,* p. 215. (Obsidian may contain spherulites.)
2. Wholly glass; solid; resinous or dull pitchy luster. *Pitchstone,* p. 218. (Pitchstone may contain spherulites.)
3. Wholly glass, but cellular or frothlike. *Pumice,* p. 218.
4. Glass, but enamel-like, and composed of small, concentric spheroids; hence fracture surfaces are botryoidal. *Perlite,* p. 218.
5. Partly glass but containing distinct, embedded crystals. *Vitrophyre,* p. 219. (The above forms are generally associated with, or grade into, felsite lavas.)
6. Glass associated with or grading into basalt; rare. *Tachylyte,* p. 220.

B. So fine grained as to be irresolvable by the unaided eye; dull or stony, not glassy.

1. Very soft: can be scratched with the fingernail.
    - *a.* Has a strong earthy or clay odor when breathed upon; if rubbed strongly between the fingers, has ultimately a smooth, greasy feeling; does not effervesce with acids; various colors. *Clay,* p. 253.
    - *b.* Friable; crumbly; soils the fingers; little or no clay odor; lively effervescence with acids; color white or light yellowish. *Chalk,* p. 261, or perhaps *marl,* p. 262. (Marl may give a distinct clay odor.)
    - *c.* General characters as in *b,* but does not effervesce with acids. *Diatomite,* p. 275.
    - *d.* Harder, more compact than *a, b,* and *c*; no clay odor; does not effervesce; composed of a mineral with a good cleavage; rarely fibrous; occurs in beds or veins. *Gypsum,* p. 270.
    - *e.* White to green, or gray; does not effervesce; no clay odor; mass has a soft, greasy feel; is foliated or shows a micaceous cleavage; folia inelastic; marks cloth. *Talc rock,* p. 318.
2. Not scratched by the nail, but easily scratched or cut with the knife.
    - *a.* Composed of excessively fine, almost imperceptible particles; dull luster and homogeneous appearance; gives clay odor when breathed on; feeble effervescence or none; has a laminated or stratified structure and usually breaks easily into chippy flakes; generally gray, but may be red, yellow, brown, bluish, or black. *Shale,* p. 253.
    - *b.* Clay odor absent or feeble; brisk effervescence with acid; white streak; commonly gray, but also white to brown or black. *Limestone,* p. 258.
    - *c.* As in *b,* but feeble or no effervescence in cold acid, but becomes brisk when rock is powdered or when the acid is heated; generally white, yellowish, or pale brown. *Dolomite,* p. 265.
    - *d.* Pale to dark green or black, rarely reddish; soapy or greasy feel; translucent on thin edges; waxy or oily appearing; subconchoidal or splintery fracture; no effervescence. *Serpentinite,* p. 330.
3. Not scratched or cut with the knife; scratches glass; does not effervesce with acid.
    - *a.* Various colors: white to red or purple, brown to dark gray; may give a clay odor; frequently shows banded flow structure. *Felsite,* p. 202.
    - *b.* Very hard: any corner or angle scratches feldspar; no clay odor; scratches steel readily; pale colors to brown or black; pronounced conchoidal fracture; glimmering horny appearance. A siliceous rock; *chert* or *flint* (dark gray to black), p. 273, or perhaps the rhyolite variety of *felsite,* p. 161.

- *c.* Not so hard as *a* and *b*; does not scratch feldspar; color black, dark gray, or green; heavy; may have a cellular or slaggy structure; may contain amygdules. *Basalt,* p. 209.

C. Distinctly granular to the unaided eye; grains wholly or partly determinable.

1. Easily scratched with the knife.
   - *a.* Effervesces briskly with cold acid. *Limestone,* p. 258, or more probably *calcite marble,* p. 327.
   - *b.* Effervesces with cold acid only when the powdered rock is treated. *Dolomite marble,* p. 265.
   - *c.* Does not effervesce with acid; probably granular crystalline. *Gypsum,* p. 270, or *anhydrite,* p. 271.
   - *d.* Soluble in water; distinct saline taste. *Rock salt,* p. 271.
2. Hard; cannot be scratched with the knife, or is scratched with difficulty. Silicate rocks.
   - *a.* Composed of grains of approximately equal size; i.e., *equigranular,* like granite; massive. See p. 166.
     - i. Mainly or wholly composed of quartz and feldspar. *Granite,* p. 166. See also *aplite,* p. 175.
     - ii. Mainly or wholly composed of feldspar but without quartz. *Syenite,* p. 180. See also *nepheline syenite,* p. 182, and *anorthosite,* p. 191.
     - iii. Composed of feldspar and a dark ferromagnesian mineral; the latter is less than or equals the feldspar. *Diorite,* p. 184. The dark mineral equals or exceeds the feldspar. *Gabbro,* p. 187.
     - iv. Composed entirely, or almost entirely, of ferromagnesian minerals; generally heavy and dark green to black (if yellowish, *dunite*). *Peridotite, pyroxenite,* etc., p. 194.
     - v. Composed of grains of quartz; scratches glass or feldspar readily. *Sandstone,* p. 249, or *quartzite,* p. 312.
     - vi. Massive silicate rocks produced by metamorphism occur in this division, but much less commonly than in the preceding. There are several varieties, depending on the particular mineral, or minerals. *Epidote rock* and *garnet rock* are examples. See *contact metamorphism,* pp. 149–153, and *carbonate-silicate rocks,* p. 327.
   - *b.* Has large, distinct crystals embedded in a finer-grained groundmass; *i.e.,* a *porphyry.* See p. 198.
     - i. Phenocrysts of feldspar and quartz with some of a ferromagnesian mineral (generally biotite) in a phanerocrystalline groundmass of feldspar and quartz. *Granite porphyry,* p. 199.
     - ii. Phenocrysts of orthoclase (and generally of a ferromagnesian mineral) in a phanerocrystalline groundmass of predominant feldspar. *Syenite porphyry,* p. 199.
     - iii. Phenocrysts of ferromagnesian minerals, or feldspar, or both, in a phanerocrystalline groundmass of feldspar and ferromagnesian minerals; feldspar phenocrysts are striated. *Diorite porphyry,* p. 201.
     - iv. Phenocrysts of quartz, or feldspar, or both, and generally of ferromagnesian minerals, in a predominant groundmass of light color and aphanitic aspect. *Felsite porphyry,* p. 203.
     - v. Phenocrysts of feldspar, or of a ferromagnesian mineral, or both, in an aphanitic, dark to black, and heavy groundmass. *Melaphyre,* p. 209.

D. *Has a distinctly foliated structure—gneissose,* schistose, or slaty.*

1. Medium to coarse-grained and roughly foliated. *Gneiss,* p. 302.
2. Consists mainly or largely of mica; much quartz is generally present; may contain porphyroblastic crystals of dark-red to black garnet, more rarely staurolite, kyanite, etc. *Mica schist,* p. 309.
3. Medium green, dark green, or black; consists mostly of a felted or matted mass of small to minute, bladed or needle-like crystals arranged mostly in one general direction, which accentuates the schistose cleavage. Other minerals, such as garnet, may be present. *Hornblende schist* or *amphibolite,* p. 321.
4. Extremely fine grained and fissile, splitting easily into thin, more or less tough, ringing slabs; usually dark gray, or green to black, but may show other colors. *Slate,* p. 314.
5. Fissile; pronounced glossy or silky luster on the foliation surfaces. *Phyllite,* p. 317. (Some phyllites contain large crystals of staurolite, garnet, biotite, etc.)
6. Soft, greasy feel; marks cloth; easily scratched with the fingernail; whitish to light gray, or green. *Talc schist,* p. 318.
7. Smooth feel; soft; green to dark green; glimmering luster. *Chlorite schist,* p. 319.

E. *Has a clearly fragmental composition; is composed of fragments or pebbles of other rocks, or of smaller pieces of angular or rounded mineral fragments; if the rock is fine grained it may show evidence of stratification.*

1. The pebbles exceed 2 millimeters in diameter and are rounded; quartz pebbles are common; the pebbles are embedded in more or less of a cement. *Conglomerate,* p. 245.
2. The pebbles are angular in shape. *Breccia,* p. 247.
3. Composed of various-sized angular fragments of volcanic rocks, such as felsite and felsite porphyry, or bits of pumice, or cellular lava, or of volcanic bombs, etc., mixed with fine compacted material (volcanic ash). *Volcanic tuff* and *breccia,* p. 224.
4. Composed of angular, less commonly of rounded grains, smaller than 2 millimeters in diameter; most or all consist of quartz, and scratch feldspar. Some cement is generally present, which, if the rock is light colored, tends to effervesce with acid (calcium carbonate); if red or brown, does not. *Sandstone,* p. 249.
5. As in 4, but feldspar is sufficiently prominent so that the rock simulates a granite in appearance. *Arkose,* p. 252.

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* Gneissose and gneissic are synonymous.

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(Rock names are printed in italics.)